

# The Effect of Water-Column Resonance on the Spectra of Secondary Microseism P-waves

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## Key Points:

- Secondary Microseisms
- Body Waves
- Source Spectra

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**Abstract**

We compile and analyze a dataset of secondary microseismic P-wave spectra that were observed by North American seismic arrays. Two distinct frequency bands, 0.13–0.15Hz and 0.19–0.21Hz, with enhanced P-wave energy characterize the dataset. Cluster analysis allows to classify the spectra and to associate typical spectral shapes with geographical regions: Low frequency dominated spectra (0.13–0.15Hz) are mostly detected in shallower regions of the North Atlantic and the South Pacific, as well as along the Central and South American Pacific coast. High frequency dominated spectra (0.19–0.21Hz) are mostly detected in deeper regions of the North-Western Pacific and the South Pacific. For a selected subset of high quality sources, we compute synthetic spectra from an ocean wave hindcast. These synthetic spectra are able to reproduce amplitude and shape of the observed spectra, but only if P-wave resonance in the water column at the source site is included in the model. Our datasets therefore indicate that the spectral peaks at 0.13-0.15Hz and 0.19-0.21 Hz correspond to the first and second harmonics of P-wave resonance in the water column that occur in shallower ocean depths (<3000m) and in the deep ocean (~5000m), respectively. This article demonstrates the important effect of water column resonance on the amplitude and frequency of P-waves that are generated by secondary microseisms, and that the amplitude of high quality sources can be predicted from ocean wave hindcasts within a factor of 0.4 – 6.

**1 Introduction**

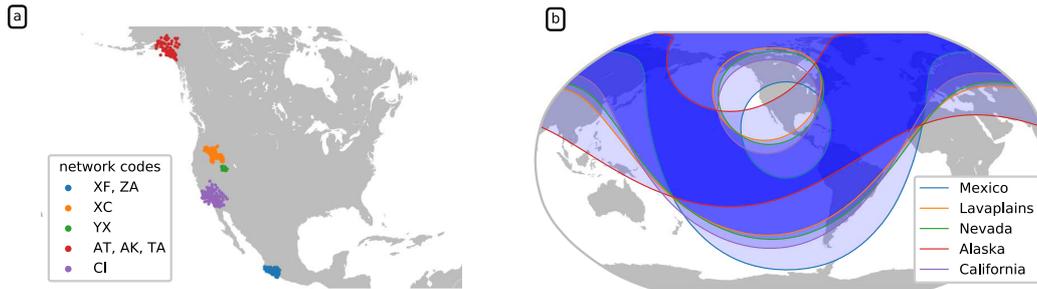
Since the first discoveries of secondary microseismic signals [e.g. *Bertelli*, 1872], numerous studies have contributed to the understanding of their origins and composition [see e.g. historical review by *Ebeling*, 2012].

Most seismic energy in the secondary microseism frequency band from 0.1–0.3Hz is associated with the interaction of opposing ocean waves [*Longuet-Higgins*, 1950; *Hasselmann*, 1963]. Rayleigh waves as well as local, regional and teleseismic P-waves have been identified at these frequencies that likely originate from this mechanism (for Rayleigh waves see e.g. [*Friedrich et al.*, 1998; *Schulte-Pelkum et al.*, 2004; *Stehly et al.*, 2006; *Stutzmann et al.*, 2009; *Schimmel et al.*, 2011], for body waves e.g. [*Haubrich and McCamy*, 1969; *Toksöz and Lacoss*, 1968; *Gerstoft et al.*, 2006, 2008; *Koper et al.*, 2009, 2010; *Landès et al.*, 2010; *Zhang et al.*, 2010a; *Obrebski et al.*, 2013]). Love waves [*Nishida et al.*, 2008; *Hadziioannou et al.*, 2012; *Tanimoto et al.*, 2015] and horizontally polarized S-waves [*Liu et al.*, 2016; *Nishida and Takagi*, 2016] have also been reported but their generation mechanism remains unclear.

Analysis of surface or body waves has robustly identified different origins of ocean wave interactions, such as storms, coastal reflections or opposing swell systems [e.g. *Gerstoft et al.*, 2006, 2008; *Ardhuin et al.*, 2011; *Stutzmann et al.*, 2012; *Obrebski et al.*, 2013]. In addition, amplification of surface or body waves due to resonance at specific water depths has been associated with the preferential observation of microseismic sources at certain locations [e.g. *Kedar et al.*, 2008; *Ardhuin et al.*, 2011; *Stutzmann et al.*, 2012; *Euler et al.*, 2014; *Gualtieri et al.*, 2014].

On the basis of the theoretical studies of seismic wave excitation due to ocean wave interactions [*Longuet-Higgins*, 1950; *Hasselmann*, 1963], quantitative modeling of surface wave energy spectra from an ocean wave hindcast has been shown to predict recorded noise spectra in the secondary microseism band [*Kedar et al.*, 2008; *Ardhuin et al.*, 2011; *Stutzmann et al.*, 2012]. Unfortunately, the source region is not readily accessible because surface waves provide access to the azimuth but not to the distance between a seismic station (or array) and the source.

Forty years ago, [*Vinnik*, 1973] showed the first modeling of P-waves generated by secondary microseism sources, but he neglected resonance effects in the ocean water column. Recently, *Ardhuin and Herbers* [2013], *Farra et al.* [2016] and *Nishida and Takagi* [2016]



**Figure 1.** (a) Seismic arrays in North America that have been used to gather continuous data for a combined timespan of five years. We use the following terms for these seismic networks: ‘Alaska’ refers to AK, AT, TA (2015), ‘California’ to CI (2015), ‘Mexico’ to XF, ZA (2006), ‘LavaPlains’ to XC (2008) and ‘Nevada’ to YX (2009). (b) Array coverage. We study P-waves in the distance range  $30^\circ - 90^\circ$  around each array. The intensity of blue color indicates how many arrays see a region.

proposed more accurate modeling of the energy spectra of secondary microseism P-waves. These advances allow us to perform an assessment of recorded P-wave spectra, with the major advantage with respect to surface waves that their origin can be localized using standard array processing techniques. This allows for a direct comparison with position, frequency and energy of sources that are predicted from an ocean wave model.

In this study, we systematically extract secondary microseism P-wave spectra from five years of seismic data recorded by North American arrays using beamforming techniques. We compute corresponding synthetic P-wave spectra from ocean wave hindcasts [Ardhuin *et al.*, 2011] using the formalism developed by Farra *et al.* [2016]. We statistically assess the accuracy of the modeled spectra by comparing them with observations in terms of spectral shape and amplitude. We show that secondary microseism P-wave energy is preferentially concentrated around two distinct frequency peaks at 0.14Hz and 0.20Hz, and that these spectral peaks are associated with P-waves from particular geographical regions. We explain these particular spectral shapes as a consequence of water column resonance at different ocean depths. Our analysis shows that water column resonance is essential to understand the origin, amplitude and spectral shape of observed P-waves from secondary microseisms.

## 2 Data and Methodology

### 2.1 Data

Our dataset consists of continuous vertical-component ground-motion data sampled at 1Hz, recorded over a combined timespan of five years by the five different North American arrays shown in Fig. 1a. We obtain data from the IRIS data management center for the year 2015 from the AK, AT and TA networks (hereinafter ‘Alaska’), for the year 2015 from network CI (‘California’), for the year 2008 from network XC (‘Lavaplains’), for the year 2011 from network YX (‘Nevada’), for the year 2006 from networks XF and ZA (‘Mexico’) (see Acknowledgements for detailed references). Note that not all days are available for all stations of these arrays during these years.

We deconvolve the instrument response function from the raw recordings, and end up with continuous ground displacement time series with flat frequency response in the secondary microseismic band (0.05–0.30Hz). Considering that we only use P-waves in the distance range from about 30 to 90 degrees from the arrays, our dataset covers the area shown in Fig. 1b, where the blue color intensity indicates the number of arrays that can see a specific

position. P-waves from distances shorter than 30 degrees or further than 90 degrees are not considered in this study because their amplitudes are difficult to model within our theoretical framework. The spatial coverage that we obtain with the arrays used in this study allows to make independent observations of the same source regions ensuring robust observations.

## 2.2 Beamforming

The central quantity of this article is the day averaged power spectral density  $B(f, \mathbf{s})$  (*beam-PSD*) at frequency  $f$  and horizontal slowness  $\mathbf{s}$  of the vertical component ground displacement recorded by a seismic array.  $B(f, \mathbf{s})$  can be seen as the plane wave energy that arrives at the array with a certain frequency and horizontal slowness from a specific direction. Such plane waves can be potentially associated with the wavefront of a distant point source. In case of a body wave with secondary microseismic origin this point source is localized at the intersection of the wave's ray-path and the surface of the Earth. We compute the beam-PSD using a standard beamforming approach that we summarize in the following paragraphs [for an extensive introduction see e.g. *Rost and Thomas, 2002*].

Initially a subwindow beam-PSD  $B_w(f, \mathbf{s})$  is computed in time windows of length 128s from the traces that are recorded by the stations in the array. In each window, traces are discarded whose standard deviation is larger than 2 times the median standard deviation of all traces of the array. This selection process removes stations with potentially corrupted data, such that we end up with  $N$  traces  $S_j(t)$  that have consistent amplitudes, where  $j$  indicates the station index within the array.

$B_w(f, \mathbf{s})$  is then computed on a square Cartesian slowness grid in the range  $-0.1$ – $0.1$ s/km, and in the frequency band from 0.08–0.30Hz that comprises the slowness range of P-waves as well as the frequency band of the secondary microseismic peak.

To compute  $B_w(f, \mathbf{s})$ , the Fourier spectra  $S_j(f)$  are phase aligned by multiplication with  $e^{-i2\pi f \mathbf{s} \cdot (\mathbf{x}_j - \mathbf{x}_c)}$  at each grid point  $\mathbf{s}$ . Here  $\mathbf{x}_j$  and  $\mathbf{x}_c$  are the positions of station  $j$  and the array center, respectively. The mean over these phase aligned Fourier spectra approximates the Fourier spectrum of the plane waves that arrive at the array center with slowness  $\mathbf{s}$ .  $B_w(f, \mathbf{s})$  is their power spectral density with unit [ $m^2/Hz$ ]:

$$B_w(f, \mathbf{s}) = \frac{1}{N^2} \left| \sum_{j=1}^N S_j(f) e^{-i2\pi f \mathbf{s} \cdot (\mathbf{x}_j - \mathbf{x}_c)} \right|^2 \quad (1)$$

$B_w(f, \mathbf{s})$ , as calculated in Eq. 1, does not only depend on the phase of the arriving waves, but also on their amplitude. Therefore, strong incoherent energy can potentially lead to larger values than weak coherently arriving plane waves. We introduce here an extra weight factor  $w(f, \mathbf{s})$ , which measures the coherency of the incoming wavefield independently of its amplitude. We use the phase stack to define this coherency factor [*Schimmel and Paulssen, 1997; Schimmel and Gallart, 2007*]:

$$w(f, \mathbf{s}) = \frac{1}{N^2} \left| \sum_{j=1}^N \frac{S_j(f)}{|S_j(f)|} e^{-i2\pi f \mathbf{s} \cdot (\mathbf{x}_j - \mathbf{x}_c)} \right|^2 \quad (2)$$

The weight factor  $w(f, \mathbf{s})$  is one for perfectly coherent plane waves and drops to a small value otherwise. The final *phase-weighted* beam-PSD reduces incoherently arriving energy and is constructed as:

$$B_{pw}(f, \mathbf{s}) = w(f, \mathbf{s}) B_w(f, \mathbf{s}) \quad (3)$$

Importantly, the phase-weighted beam-PSD of a coherently arriving plane wave is unchanged compared to the unweighted beam-PSD. Strong signals that are not coherent over the whole array are suppressed by this procedure. Such signals can have high amplitude and contaminate the beam-PSD. Because the microseismic body wave signals are not perfectly coherent plane waves, the observed beam-PSD might be slightly biased towards lower amplitudes. In

our experience the amplitude bias that is introduced with the phase-weight is only on the order of few percent. At the same time, it increases the slowness resolution of the beam and removes incoherent sources.

The beam-PSD can be seen as a collection of spectra (PSDs), i.e. one PSD at each slowness. Most of the time, the PSDs that are examined in this article have a single spectral peak. We will see that this is not always but most of the time the case for secondary microseisms [see also *Neale et al.*, 2017]. We can therefore use two simple parameters to describe the PSD: the PSD's *amplitude* refers to the maximum of the PSD, and the PSD *frequency* or *dominant frequency* refers to the frequency of the PSD maximum. This representation allows to use a 2D image as a compact visualization of the beam-PSD, or more specifically of the beam-PSD spectral peak: color lightness is used to show the beam-PSD amplitude, and color hue for the beam-PSD frequency [for more information on the 2D colormaps see *Meschede and Stutzmann*, 2016].

The dataset is reduced further by averaging the 128s subwindow beam-PSDs  $B_{pw}$  over full days, i.e. from 00h–24h. Before averaging, however, we remove subwindows that are potential outliers, i.e. subwindows that contain a lot more energy than the other subwindows of the day and that is unlikely to originate from secondary microseisms. To this end, we compute a single intensity value for each subwindow from  $B_{pw}$ , defined as the sum of the beam-PSD amplitudes of all slownesses (from  $-0.1\text{s/km} - 0.1\text{s/km}$ ). A subwindow is considered valid if the intensity value is smaller than 2 times and greater than 1/1000 times the median intensity value of the day. Only valid subwindows are used for the daily average.

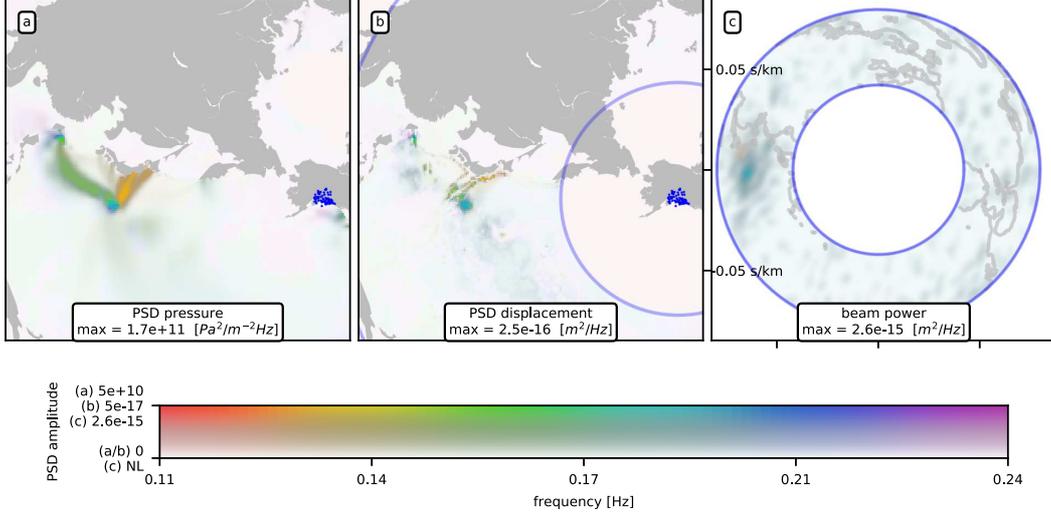
This selection procedure effectively removes (smaller to intermediate) earthquakes and other short duration outliers that are for example caused by local noise at the array. If most subwindows during a day are noisy due to larger earthquakes or other regional noise, the median intensity value rises and the selection criterion does not discard this day. We therefore end up with some noisy beam-PSDs in our dataset. The day average of the observed beam-PSD is denoted  $B(f, s)$  and from now on just called *observed beam-PSD*.

Typical beam-PSD amplitudes are high only around the few slowness values where plane waves add coherently. The beam-PSD amplitude at slowness values away from these peaks corresponds to random fluctuations because here the plane waves add incoherently in addition to other random noise. The median of the beam-PSD amplitudes (from  $-0.1 - 0.1\text{s/km}$ ) is a good measure of the amplitude level of these random fluctuations. The median therefore defines a reasonable threshold below which signals cannot be detected (from now on called noise level of the beam). To enhance the visual perception of the signals of interest, white color indicates in some figures amplitudes below the noise level of the beam rather than zero (indicated by *NL* in plot colorbar labels).

To stabilize the beam-PSD amplitudes, we smooth it over a narrow frequency band of about 0.04Hz (Hanning window with width 0.08Hz) before computing the beam-PSD maxima at each slowness. Very narrow-band sources with bandwidths of less than 0.04Hz are attenuated by this procedure. To compute the dominant frequency of the PSD, we use the original unsmoothed spectra.

### 2.3 Synthetic Beams

We use the ray theoretical approach proposed by *Farra et al.* [2016] to compute the *synthetic beam-PSD*  $B_{syn}(f, s)$  from global microseism pressure-PSD models that are derived from the ocean wave hindcast WAVEWATCH III [*Tolman et al.*, 2009; *Ardhuin et al.*, 2011; *Rascole and Ardhuin*, 2013] and openly distributed. Modeling consists in 3 essential steps, described in detail in [*Farra et al.*, 2016] that are summarized in Fig. 2 and described hereafter:



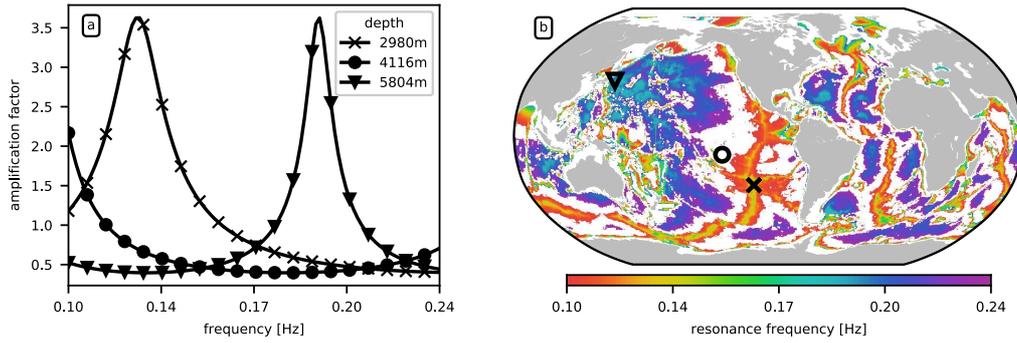
**Figure 2.** Example of synthetic beam construction for the Alaska array on day 2015–08–23. (a) Pressure-PSD (saturated colors) that is obtained from model R. (b) Displacement-PSD (saturated colors) at the array center due to the pressure-PSD shown in (a) backprojected on map. The minimum of the colormap (white color) is set to zero in (a) and (b). The colormap saturates the PSD amplitudes at (a)  $5 \times 10^{10} \text{ Pa}^2/\text{m}^2\text{Hz}$ , and at (b)  $5 \times 10^{-17} \text{ m}^2/\text{Hz}$ . The displacement-PSD amplitudes that are greater than  $1 \times 10^{-17} \text{ m}^2/\text{Hz}$  are taken into account to compute the synthetic beam-PSD shown in (c). In (c) the colormap minimum (white color) is set to the noise level (NL) of the beam that is about 1/7 of the strongest beam-PSD amplitude. In (a)-(c) blue circles delimit the P-wave distance range that we consider ( $30^\circ - 90^\circ$ ).

First, we compute  $F_p(f, \mathbf{x})$ , the power spectral density of the pressure fluctuations (*pressure PSD*) that are generated by wave-wave interactions, i.e. the power per frequency  $f$  and horizontal wave vector  $\mathbf{k}$  with unit  $[\text{Pa}^2/(\text{m}^2\text{Hz})]$  at position  $\mathbf{x}$ . Pressure fluctuations at frequency  $f$  and wave vector  $\mathbf{k}$  are generated by interacting ocean waves with frequency  $f_1, f_2$  and wave vectors  $\mathbf{k}_1, \mathbf{k}_2$ , where  $f = f_1 + f_2$  and  $\mathbf{k} = \mathbf{k}_1 + \mathbf{k}_2$ . Only nearly opposing ocean waves with  $\mathbf{k}_1 \approx -\mathbf{k}_2$  and  $f_1 \approx f_2 \approx f/2$  generate the pressure fluctuations that are necessary to excite seismic waves. Based on statistical hypotheses about the ocean wave height spectrum,  $F_p$  is independent of  $\mathbf{k}$  in the range of wave vectors that correspond to seismic P-waves, and the pressure fluctuations are spatially uncorrelated on scales that are larger than the wavelength of seismic P-waves in water [Hasselmann, 1963].

With these assumptions,  $F_p$  is computed from the ocean wave height energy spectrum  $E(\mathbf{x}, f/2)$  and its directional distribution function  $M(\mathbf{x}, f/2, \phi)$  with azimuth  $\phi$ .  $E$  and  $M$  are provided by the ocean wave hindcast [for details see Tolman *et al.*, 2009]. Computing  $F_p$  from  $E$  and  $M$  consists essentially in an integration of the ocean wave height energy spectrum over opposite azimuth  $\phi$  [Hasselmann, 1963; Ardhuin *et al.*, 2011]:

$$F_p(\mathbf{x}, f) = [2\pi]^2 [\rho_w g]^2 f E^2(f/2) \times \int_0^\pi M(f/2, \phi) M(f/2, \phi + \pi) d\phi \quad (4)$$

Here  $\rho_w$  is the water density and  $g$  the gravitational acceleration at the surface of the Earth. The resulting pressure PSD field has twice the frequency than the interacting waves. Ocean waves without opposing counterpart do not generate a pressure perturbation that can be observed seismically.



**Figure 3.** (a) P-wave amplitude amplification due to water-column resonance as a function of frequency for three different ocean depths, computed for slowness 0.05 s/km in a simple two layer model. (b) Map of the corresponding resonance frequencies in the frequency band of interest (0.10–0.24Hz) and at slowness 0.05 s/km. The position of the three examples from (a) are indicated with the corresponding symbols in the map.

Fig. 2a shows the pressure-PSD field for day 2015-08-23, shown from the perspective of the Alaska array. As explained in the previous section, the pressure PSD is characterized at each position by the PSD maximum (*pressure-PSD amplitude*), shown by the color lightness, and the frequency of the PSD maximum (*pressure-PSD frequency*), shown by the color hue.

In Fig 2a, two strong pressure sources with a dominant frequency of about 0.20Hz (turquoise color) can be seen that are generated by the two typhoons Goni off the coast of Taiwan and Atsani off the coast of Japan. Furthermore, there are strong wave interactions in between both typhoons that generate pressure sources with a dominant frequency of approximately 0.16Hz (green), and significant sources due to ocean wave coastal reflections around 0.13Hz (orange/yellow).

Fig. 2b shows the vertical ground displacement-PSD amplitude, backprojected on a map, that a seismic station located at the array center would record due to the pressure-PSD shown in Fig. 2a. Propagation effects, such as geometrical spreading and attenuation, are computed here using a ray tracing formalism [Farra *et al.*, 2016] in the model IASP91 [Kennett and Engdahl, 1991] combined with an attenuation model of Montagner and Kennett [1996].

At a particular position or slowness, these propagation effects transform the pressure-PSD in a smooth and systematic way. On the other hand, water column resonance at the source-site can sharply amplify the PSD of P-waves that are radiated from a pressure fluctuation at the ocean surface in the vicinity of a particular frequency. Resonance occurs if a P-wave that is reflected from the ocean bottom and back from the free surface adds constructively to the wavefield. Multiple reflections within the water column can then lead to strong amplification.

An estimate of the resonance frequencies can be obtained in the case of vertically traveling P-waves. In this case, the reflected waves accumulate phase corresponding to the two way traveltime in the ocean with depth  $h$  and an additional phase shift of  $\pi$  due to the free surface reflection. Constructive interference takes place when the total accumulated phase corresponds to  $2\pi n$ , where  $n$  is a positive integer number. With wavelength  $\lambda$  in the water, this can be written as

$$2\pi(2h/\lambda) + \pi = 2\pi n \quad (5)$$

or alternatively as

$$h/\lambda = (2n - 1)/4 \quad (6)$$

Resonance therefore occurs approximately when the ocean depth corresponds to  $1/4, 3/4, 5/4, \dots$  of the P-wave wavelength. Assuming a P-wave water velocity of  $\alpha_{\text{water}} = 1500\text{m/s}$ , ocean depths of 6000m generate resonances around frequencies 0.06Hz, 0.19Hz, 0.31Hz, ... , ocean depths of 4000m generate resonances at 0.09Hz, 0.28Hz, ... , and ocean depths of 2500m at 0.15Hz, 0.45Hz, ... . Only few of these resonance peaks are in the frequency band of interest where the pressure-PSD  $F_p$  has significant energy (0.10–0.24Hz). In particular, we deal with the resonance peaks for  $n = 1$  (0.15Hz) at ocean depths of 2500m depth and  $n = 2$  (0.19Hz) at ocean depths of 6000m. Intermediate frequencies (0.16–0.18Hz) are amplified by shallower depths than 2500m through the  $n = 1$  resonance or larger depths than 6000m through the  $n = 2$  resonance.

Non-vertical P-waves can be taken into account replacing the ocean wave height  $h$  by an effective height  $h_\theta = h/\cos(\theta)$ , where  $\theta$  is the angle between ray path direction in the water and the vertical axis. Typical values of  $\theta$  for P-waves are  $3.5^\circ - 6.8^\circ$  in the  $30^\circ - 90^\circ$  distance range, which does not change the resonance frequency much compared to the vertical case [see *Gualtieri et al.*, 2014].

To compute the magnitude of the amplification factor as a function of frequency, the sum of all P-waves, that are multiply reflected and then transmitted in the solid Earth, can be evaluated as a geometric series, as described in detail in *Gualtieri et al.* [2014]. The magnitude of the amplification factor depends on the P-wave reflection and transmission coefficients that we compute using a simple 1D model with a water layer over a solid halfspace. The 1D model is defined with the following structural parameters:

$$\rho_{\text{solid}} = 2500\text{kg/m}^3, \rho_{\text{water}} = 1000\text{kg/m}^3, \beta_{\text{solid}} = 3200\text{m/s}, \alpha_{\text{solid}} = 5540\text{m/s}, \alpha_{\text{water}} = 1500\text{m/s}.$$

Fig. 3a shows the P-wave amplification factor for slowness 0.05s/km, computed for three different water depths in the spectral range of interest (0.10–0.24Hz). At depths characteristic of mid-ocean ridges (2980m in Fig. 3a) it has a broad resonance peak around 0.13Hz corresponding to  $n = 1$  in Eq. 6. At intermediate depths (4116m in Fig. 3a) the amplification factor has no resonance peak in the frequency range of interest because the  $n = 1$  resonance frequency is too low and the  $n = 2$  resonance frequency too high. The amplification factor at deep ocean depths (5804m in Fig. 3a) has a sharp high frequency resonance peak around 0.19Hz, corresponding to  $n = 2$  in Eq. 6. The 0.13Hz resonance peak is broader than the 0.19Hz peak because the accumulated phase of the reflected P-wave before interference is proportional to the product of frequency  $f$  and ocean depth  $h$ . Therefore the width of the resonance peak is inversely proportional to the ocean depth.

In Fig. 2b, the low frequency coastal reflection sources, as well as typhoon Goni close to Taiwan and the sources that connect both typhoons are strongly damped compared to Fig. 2a. In these source regions, the frequencies with significant source pressure-PSD do not correspond to the resonance frequencies of the water column.

Site amplification can modulate the recorded P-wave amplitude from a factor of 0.4 to 3.7 and the beam-PSD by the square of this factor from 0.16 to 14. Nevertheless, the variations of source pressure due to the interacting ocean waves can attain similar orders of magnitude, and it is the combination of the pressure-PSD and the amplification factor that controls the amplitude and spectral shape of the radiated P-wave.

Fig. 3b shows a map of the resonance frequencies within the frequency band of interest (0.10–0.24Hz). Different symbols indicate the locations with the amplification factor shown in Fig. 3a. There are two major resonance frequencies in the oceans: high frequency resonance (0.18–0.22Hz) in the deep oceans corresponding to  $n = 2$  (blueish color) and low frequency resonance corresponding to  $n = 1$  (0.10–0.15Hz) in shallower parts over ridges

(red and yellow color). Intermediate frequency resonance (green color) is rare, very shallow coastal regions and intermediate depth ranges do not amplify P-wave energy with frequencies in the band of interest. If ocean wave pressure sources were distributed evenly over the ocean and all frequencies, we would expect secondary microseismic P-wave energy to be predominantly concentrated in the high and some in the low frequency peaks.

Finally, the computed P-wave displacement-PSD from each grid point is convolved in slowness space with the array response. The sum of all sources whose displacement-PSD is greater than a specified threshold ( $10^{-17} m^2/Hz$ ) yields the *synthetic beam-PSD*  $B_{syn}$  shown in Fig. 2c. The synthetic beam-PSD has only a single peak with a single color in contrast to Figs. 2a-b. At most slownesses the beam-PSD amplitude corresponds to random fluctuations with a beam-PSD frequency that is the same as at the isolated peak. These random fluctuations can be seen as the noise level of the beam that is at about 1/7 of the strongest beam-PSD amplitude in this particular case. The noise level is controlled by the beam-PSD amplitude of the strongest source, the overall number of sources, the geometry of the array and our beamforming technique. In this specific case the noise that corresponds to the strongest source dominates all other sources that are weaker. Multiple sources can only be analyzed if they have comparable amplitude. We therefore focus in this article only on the strongest source of a particular day.

The spatial beam-PSD resolution, controlled by the aperture of the arrays that have been used for this study, is much lower than the spatial resolution of the ocean wave model. Therefore the synthetic beam-PSD corresponds to a smoothed version of the vertical displacement PSD that has been computed in step two.

In this article we use and compare three different models for  $B_{syn}$  that vary: (a) The effect of ocean wave coastal reflections. Coastal reflections are either included in the ocean wave hindcast as coastal reflection coefficients of 10% along continents, 20% at island, and 40% at icebergs, or completely excluded. (b) The effect of water-column resonance. The amplification factor is either computed as described above, or neglected by setting it everywhere and at all frequencies to one.

*Model R*, the most comprehensive model, includes coastal reflections and water-column resonance. *Model NR* includes water-column resonance but no coastal reflections. *Model NA* includes coastal reflections but no water-column resonance. Each of the models NR and NA switches off one of the two major factors of the modeling procedure whose influence we would like to examine.

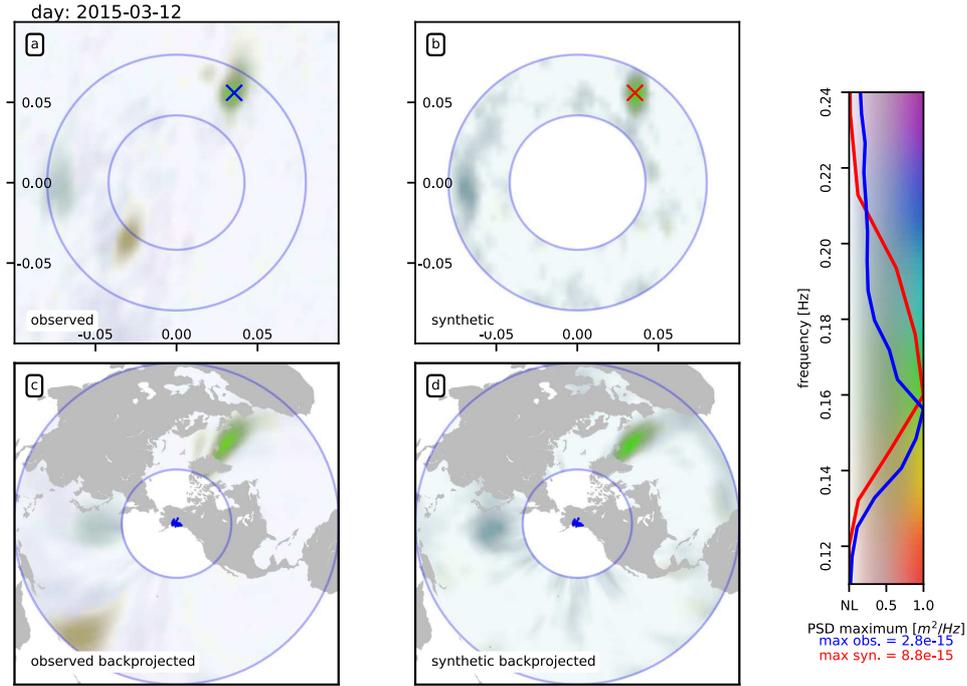
### 3 Observations

#### 3.1 Individual days

Fig. 4 illustrates the observed beam-PSD  $B$  and the synthetic beam-PSD  $B_{syn}$  of model R for day 2015–03–12. The first row shows  $B(f, s)$  (a) and model R  $B_{syn}(f, s)$  (b) in slowness domain. The beam-PSD amplitude (narrow-band smoothed) is displayed with color lightness, and the beam-PSD frequency with color hue. The second row (c/d) shows the projection of (a/b) on a map, using the P-wave slowness-distance relation.

The slowness with the highest beam-PSD amplitude (from now on *strongest-source slowness*) is indicated with a cross in Figs. 4a and b. We describe the *strongest-source PSD*, i.e. the PSD that corresponds to the strongest-source slowness, by its maximum, called *strongest-source amplitude*, and by the frequency of its maximum, called *strongest-source frequency*.

The strongest-source slowness is nearly the same for  $B$  and  $B_{syn}$ . In this particular example, P-wave energy is predominantly emitted by microseisms around Greenland. In addition, weaker sources are observed and modeled north of Japan. A third source in the South Pacific is observed but it does not appear clearly in the beam-PSD of model R.



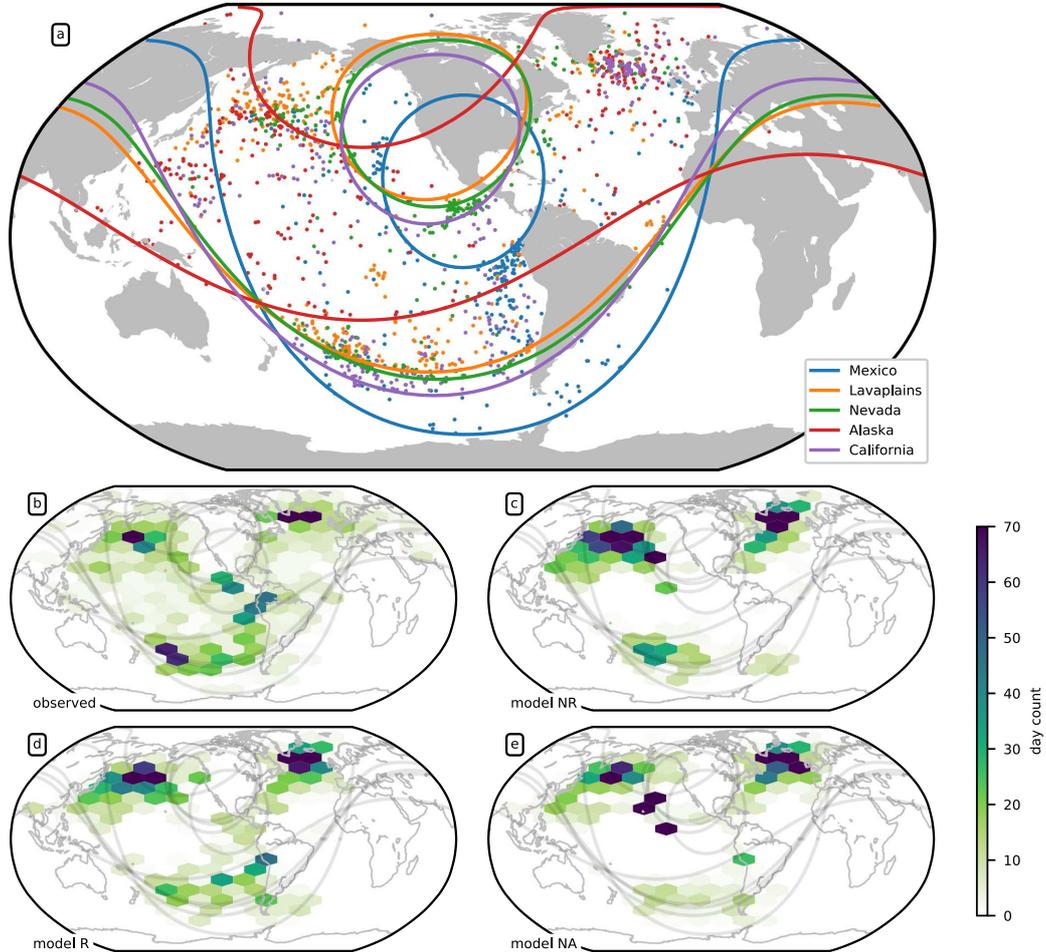
**Figure 4.** Observed (a/c) and synthetic (b/d) beam-PSD at the Alaska array on day 2015–03–12. Color hue represents the dominant frequency and color lightness the spectral maximum of the beam-PSD (narrow-band smoothed). The strongest-source with dominant frequency around 0.16Hz (green) is located south of Greenland, some higher frequency microseismic energy is generated off the Japanese coast with dominant frequency around 0.20Hz, and there is a clear observed source with low dominant frequency of 0.13Hz in the South Pacific. The red and blue spectra in the colorbar refer to the strongest source spectra from a and b that are indicated by a cross.

The observed strongest-source amplitude, corresponding to the Greenland source, is with  $2.8 \times 10^{-15} m^2/Hz$  approximately 3.1 times weaker than the predicted one with  $8.8 \times 10^{-15} m^2/Hz$ . The strongest-source PSDs are plotted in red and blue in the colorbar of Fig. 4. The observed (blue) and modeled (red) beam-PSD peak around 0.16Hz, but the observed spectral peak is sharper and better defined. The beam-PSD amplitudes of the Japanese source cannot be precisely compared from this plot but its observed and modeled dominant frequencies of about 0.20Hz (blue color) are similar. The observed South Pacific source has a low dominant frequency of approximately 0.13Hz.

Fig. 4 shows a case where observed and modeled beam-PSD have some similarities but also some differences. We observe the same principal similarities and differences using the California array on the same day. It is interesting to understand, in particular, why the South Pacific source does not appear in the synthetic beam. Inspection of the input pressure-PSD and displacement-PSD of the model (not shown) at the position of this source reveals that it does exist and that it corresponds to a tropical cyclone called Pam. However, the amplitudes of pressure-PSD and displacement-PSD in these regions are so low that this source disappears under the noise level of the much stronger and higher frequency sources in Greenland in the final synthetic beam-PSD. It is unclear, which part of the modeling process is responsible for this mismatch between observed and synthetic beam. The South Pacific source could be too weak, but also the amplitude of the Greenland source, that is clearly overpredicted in model R, raises the noise level of the synthetic beam.

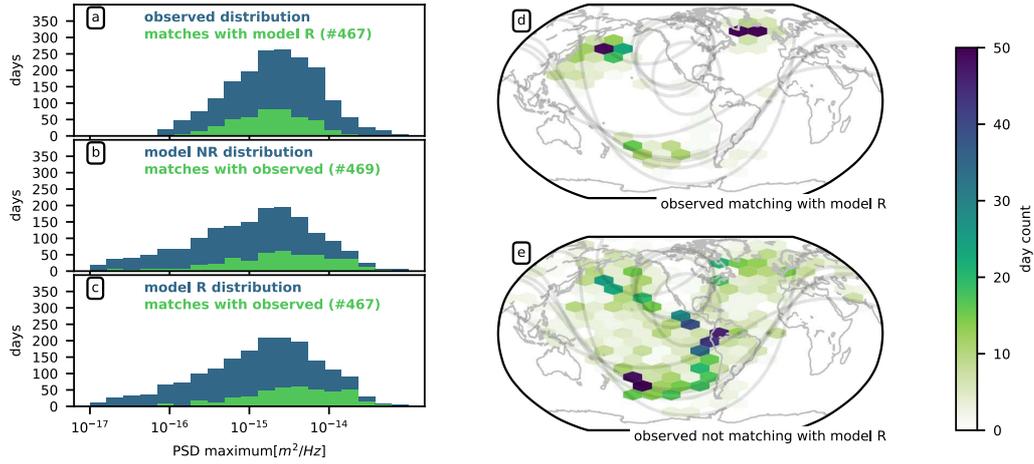
To obtain conclusions about the general quality of the modeling process, we statistically compare observations as the one shown in Fig. 4 with the modeled beam-PSD in the following sections. Because weaker sources can disappear in the beam-PSD behind the noise of significantly stronger ones, as we have seen in the previous example, we concentrate in this article on the strongest-sources only that provide the most robust measurements.

### 3.2 Strongest-Sources: Energy Distribution and Geographical Origins

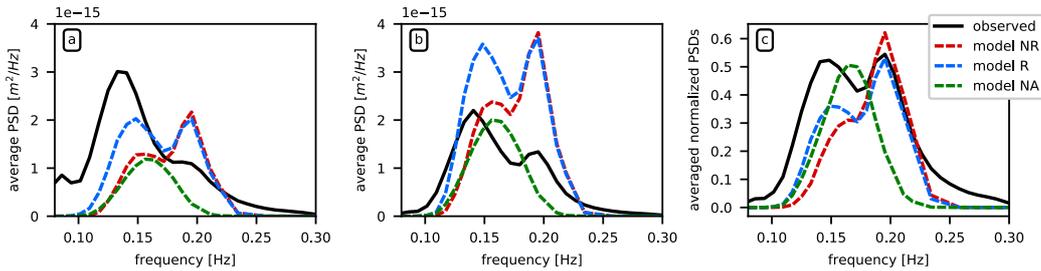


**Figure 5.** Geographical distribution of the strongest-sources. (a) shows the observed strongest-source positions colored by the array that observed them. We have added a small jitter ( $2^\circ$ ) to the positions to make clusters better visible. Observations that cluster at the observation limits of an array have to be interpreted with caution. The source density is shown for observations, model NR, model R, and model NA from (b) to (e). The observation limits of the arrays are indicated with lines from (a) and (e). The observed and modelled source density is biased by the number of arrays that can observe a particular region (see Fig. 1b). Model R fits best with the observations. Model NR and NA are missing the coastal reflections along the South American Pacific coast. Model NA has the worst overall fit with the observations.

The *strongest-source PSD* is the beam-PSD at the slowness with the highest beam-PSD amplitude, as plotted in the colorbar of Fig 4 for a particular day. From all five years and arrays we obtain 1809 strongest-source PSDs and slownesses from the observations and equally from each model. Strongest-source amplitudes and strongest-source frequencies are



**Figure 6.** Distribution of the observed (a), model NR (b) and model R (c) strongest-source amplitudes from all five years observed by the five arrays (1809 total days of observations). Sources are called matching when the observed and modelled strongest-sources slowness is similar. The matching source distribution is shown in green and non-matching sources in blue. The geographical positions of the observed sources that match with model R, shown in (d), are concentrated in three dominant source regions: south of Greenland, and in the North-West and South Pacific. The non-matching source positions, shown in (e), also concentrate in these regions but also along the coast of South America. The observation limits of the arrays are indicated in (d) and (e) with faint black lines.



**Figure 7.** Averages of observed and modeled strongest-source PSDs. (a) shows the average of all strongest-source PSDs, (b) shows the average of the matching strongest-source PSDs and (c) shows the average of the matching normalized strongest-source PSDs (i.e. their maximum is normalized to one).

extracted from the strongest-PSDs and describe the amplitude and frequency of their maxima as explained previously. These quantities constitute the observed and corresponding modeled datasets that we analyse and compare in the following sections.

Fig. 5a shows the positions of the strongest-sources colored by array that observed them. Some jitter ( $2^\circ$ ) was added to the source positions to better resolve nearby points. Importantly, this plot considers all observed sources to be secondary microseism P-waves although they could also correspond to some other source of seismic energy such as another body wave phase (e.g. PP), or leaked energy from strong surface waves. Many source regions, e.g. in the North Pacific, North Atlantic and South Pacific, are seen individually by several arrays. Others, as at the Pacific coast of South and Central America, can be predominantly seen by one of the arrays.

Some sources can be found on the continents and are certainly not secondary microseism P-waves. Other observations cluster at the P-wave observation limits of the arrays, such as around northern South America and the Pacific coast of Central America. Although such sources could be generated by secondary microseisms, they have to be examined with care because they can correspond to energy of a regional phase (for short distances), or a P-wave that is diffracted around the Earth's core (for large distances). Such signals can leak into the P-wave slowness range and are not modelled in the synthetic beam. In these two cases, the source distances would be over or underestimated, respectively. In particular the Mexico and Nevada arrays suffer from higher leakage because they are smaller.

The strongest-source density is shown for the observations, model NR, model R and model NA from Fig. 5b-e. The principal observed source regions are the North-West Pacific, the South Pacific, the North Atlantic, as well as the South and Central American Pacific coast. Only model R approximately reproduces all of these source regions. The South and Central American Pacific coast sources do not appear in model NR and are underestimated by model NA, although the latter includes coastal reflections in the same way as model R. Model NA furthermore underestimates the strongest-source distribution in the South Pacific, and diverges in general most from the observed one.

Many of the observed strongest-sources in Fig. 5 do not correspond to clean P-waves from a secondary microseism and cannot be compared to the modeled ones. The strongest sources cannot be impulsive signals such as earthquakes because they were removed in the processing. On the other hand, the strongest sources could correspond to other microseismic body wave phases such as PP, or to leaked microseismic energy from surface waves. Furthermore these signals could be generated by unknown, non-microseismic processes. Even in the case that an observed strongest-source corresponds to a secondary microseism P-wave, the corresponding modelled strongest-source of the same day could be in a completely different region of the Earth.

A comparison can only be made between observed and modeled sources that correspond to each other, i.e. that have similar slowness and therefore come from the same position on the same day. Also, sources with a matching modelled counterpart are likely P-waves from a secondary microseisms: It is very unlikely (not impossible) that a signal other than a secondary microseism P-wave has similar slowness on the same day as the strongest-source P-wave derived from the wind-derived secondary microseism model. Selecting strongest-sources with matching modelled counterpart therefore removes observations of other body wave phases, leaked energy, or signals of non-microseismic origin. Importantly, this selection mechanism does not imply that we select any a-priori strongest-source PSD. On the other hand, we do favour source regions that appear in the models, which is necessary for a meaningful spectral comparison.

We therefore call days 'matching' with a particular model when the observed and modeled strongest-source slownesses on this day are closer than 0.01 s/km (corresponding to a distance of approximately 1700km). Otherwise days are called 'non-matching'. The matching days provide us with 469 (model NR), 467 (model R) and 318 (model NA) observed and modeled strongest-source PSDs, which correspond to 26% for models NR and R, and 18% for model NA of all 1809 days.

Note that an observed strongest-source can be accurately modeled but still be classified as 'non-matching'. This happens for example if another stronger source exists in the model elsewhere which would be selected as strongest-source. 'Matching' or 'non-matching' in itself is therefore not a good criterion of the ocean wave model accuracy but a strict data selection procedure that allows us to extract source spectra from the full dataset that can be directly compared. It is not our objective to reach the highest number of matching sources but a set of high quality sources that are very likely to be secondary microseism P-waves, and that are the least perturbed by other stronger signals. The dataset of observed strongest-

sources matching with model R is shown in detail in Table S1 in the supporting information material.

From Table S1, it can be seen that the strongest-source spectra have most of the time a single peak. The few more complex spectra that we observe can be the consequence of the limited spatial array resolution that mixes P-wave energy from a large region or also of perturbations by other signals. This is in particular the case for observations from the smaller Mexico and Nevada arrays. During the year 2015, we have many independent observations of the same source from the two North American arrays in ‘California’ and ‘Alaska’ (days: 2015-01-05, 2015-01-07, ...). Comparing the spectra of the same source observed by these two arrays (e.g. day 2015-01-05), confirms that their strongest-source frequencies are in general close, and therefore a property of the source rather than a consequence of the observation with a particular array. Some variations in the spectral shape (e.g. day 2015-01-08) give an idea of the differences due to observations from different arrays. Many of these differences can be attributed to contamination by strong local noise that influences one array more than another.

Fig. 6 shows the full distribution (blue) of the strongest-source amplitudes of all 1809 days corresponding to the observations (a), model NR (b) and model R (c). In each distribution, the portion of strongest-sources matching with either model R (a), and the observations (b) and (c) is shown in green.

In contrast to the observed distribution (a), the modeled distributions (b/c) have a tail that extends into low amplitudes. This is not surprising, because the observed strongest-source PSDs are to some extent biased by noise. The observed distribution (a) is concentrated more strongly around its center but has a faint high amplitude tail that does not appear in (b) and only a little bit in (c).

It is surprising, in Fig. 6a, that the observed strongest-sources matching with model R (green) have approximately the same distribution as the non-matching ones (blue). This means that weak observed strongest-sources are as likely to have a modeled match as the strong ones. This is different for the modeled distributions (Fig. 6b/c): here, clearly the matching strongest-sources (green) tend to have higher beam-PSD amplitudes than the non-matching ones (blue). Therefore strongest-sources that are strong in the model are more likely to be observed, as we would expect.

Fig. 6 shows maps of the *observed* geographical positions of the strongest-sources that match (d) and do not match (e) with model R (corresponding to the distribution in Fig. 6a). The matching strongest-sources focus in three dominant source regions south of Greenland, in the North-West Pacific and in the South Pacific.

The geographical positions of the observed strongest-sources that do not match with model R are widely distributed over the oceans. Less so than the matching sources, they still clearly concentrate in some specific regions, most notably in the North-Central Pacific, along the South and Central American coastlines, and in the South Pacific. Although some of these non-matching observations are surely contaminated by noise, their non random regional distribution implies that they are to a large extent also related to secondary microseisms. In particular the strong concentration of non-matching sources at the Pacific Coast of northern South and Central America is likely caused by leakage from regional P-waves to the Mexico and Nevada arrays. These arrays are fairly small and leakage is therefore more prominent. Similarly some of the non-matching South Pacific sources might correspond to leakage from P or Pdiff phases beyond the 90° distance range.

Fig. 7 shows the average of all strongest-source PSDs (a), of the matching strongest-source PSDs (b) and of the matching normalized strongest-source PSDs (i.e. their maximum is normalized to one) (c). The different lines correspond to the observed (black), model NR (dotted red), model R (dotted blue), and model NA (dotted green) averages.

One striking feature of the averaged PSDs in Fig. 7a-c are two distinct spectral peaks that appear in observations as well as in the models that take into account water-column resonance (models NR and R). The first peak is around 0.13–0.15Hz and the second around 0.19–0.20Hz. Model NA, which does not take into account the water column resonance, has only a single broad maximum around 0.16–0.17Hz. Model NA reflects the actual ocean wave pressure-PSD, only slightly distorted by seismic propagation effects such as attenuation. The absence of the spectral peaks in model NA indicates that they are a consequence of the resonance effect, caused by the source-site amplification shown in Fig. 3.

At low frequency, the averaged strongest-source PSD peak in Fig. 7a is at about 0.13Hz in the observations, and not like in model R at 0.14Hz, in model NR at 0.15Hz, and in model NA at 0.16Hz. Another difference is that there is more low frequency energy in all observed strongest-sources than in all modeled sources, whereas the high frequency energy is weaker in the observation. The averages of all strongest-source PSDs give an idea about the strongest-source PSDs before selecting matching sources only but they are not directly comparable because they might correspond to completely different sources.

If we look at the average strongest-source PSD of the matching days only, shown in Fig. 7b, the observed excess low frequency energy vanishes. Here, the modeled amplitudes are stronger at all frequencies compared to Fig. 7a. One reason is that strongest-sources with modeled high amplitude are matching more often with observations (see Fig. 6b/c). The average observed strongest-source PSD on the other hand becomes weaker at low frequencies and slightly stronger at high frequencies when only matching sources are used. This means, that the non-matching sources are dominated more strongly by low frequency energy. It remains unclear from Fig. 7a-b whether the average strongest-source PSDs are mostly influenced by the spectral shape of the ensemble of the sources or due to a few particularly strong ones.

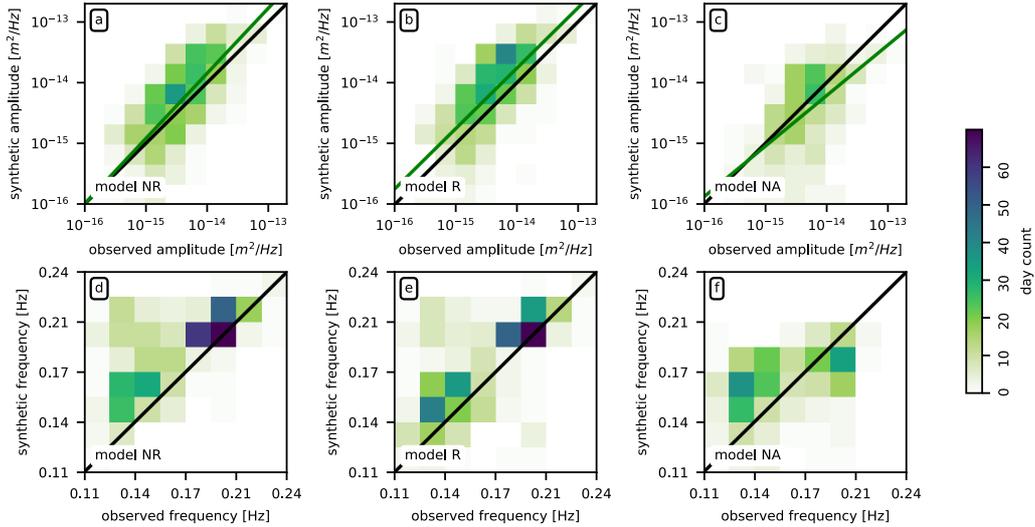
Normalizing the spectra before averaging ensures that all sources contribute in the same way to the average. The average of the matching normalized strongest-source PSDs is shown in Fig. 7c, and shows that there are as many observed low frequency and high frequency sources. The models, however, have more high frequency than low frequency sources. A comparison with the unnormalized averages in Fig. 7b shows that normalization reduces the amplitudes of the low frequency sources compared to the high frequency ones. This is a hint that the low frequency matching sources are in general more energetic than the high frequency ones in the observations and in the models.

We conclude several things from this section: secondary microseism P-wave energy concentrates in two spectral peaks, likely because of resonance effects in the source-site water column. The observed strongest-sources match in slowness with the modeled ones on about 26% of all days for models NR and R and on about 18% days for model NA. Models NR and R can reasonably well explain the double peak of the averaged matching source spectra. However, the observed low frequency resonance peak is at slightly lower frequencies than in the models. Also, the modelled matching sources are more often high frequency dominated than the observed ones, and their amplitudes are slightly higher in general. Low frequency sources are more energetic than the high frequency sources in observations and models. The site-effect amplification therefore seems important to quantitatively model observed P-wave spectra.

### 3.3 Matching Strongest-Sources: Energy and Dominant Frequency

Fig. 8 shows a comparison of the observed with modeled matching strongest-source amplitudes (row 1) and frequencies (row 2) for model NR (a/d), model R (b/e) and model NA (c/f).

The (geometric) average of the strongest-source amplitudes is accurately modeled to a factor of 1.3 (a), 1.7 (b) and 0.70 (c). The standard deviations around a linear least squares



**Figure 8.** Comparison of matching observed and modelled strongest-source amplitudes (row 1) and strongest-source frequencies (row 2) from all five years and arrays. The observed amplitudes are compared with models NR (a), model R (b) and model NA (c), and the observed frequencies with model NR (d), model R (e) and model NA (f). The lines show equality (black) and a linear least square fit (green) in log-log space.

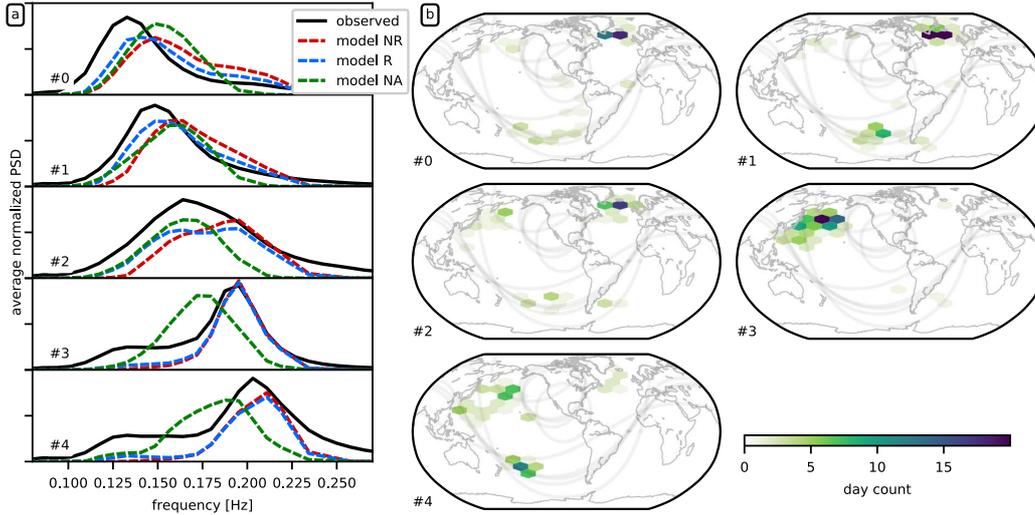
fit of the logarithmic strongest-source amplitudes correspond to factors of 3.2 (a), 3.1 (b) and 3.8 (c), respectively. The matching strongest-source amplitudes are therefore roughly predicted within a factor of 0.4-6 by the models. All models predict the strongest-source amplitudes of the matching sources reasonably well. These numbers should be seen in the perspective of ocean wave energy that can vary by multiple orders of magnitude.

Amplitudes tend to be slightly overestimated by models NR (a) and R (b). Model NR has slightly weaker amplitudes than model R, which adds energy due to coastal reflections, and therefore fits the observed amplitudes a little bit better at low frequencies. The amplitude of model NA also fits approximately the observed values but it depends on the a-priori unit amplitude that we have chosen for the site amplification factor. Note that these details can be independently seen at all arrays and years (see Fig. A1).

The strongest-source frequency shown in Figs. 8d-e indicate two principal groups of strongest-source PSDs corresponding to the two spectral peaks seen in Fig. 7. There is a group of low frequency sources with dominant frequencies around 0.14Hz, and there is a group of high frequency sources with dominant frequencies around 0.20Hz. These two groups, with a gap in between, are modeled and observed independently in Figs. 8d-e. For most sources the observed and modeled dominant frequencies belong to the same group and are very close.

The strongest-source frequencies (Fig. 8f) of model NA do not have this characteristic. Here, the synthetic dominant frequencies are distributed smoothly over a smaller interval, while the observed have the characteristic double peak.

Sometimes, model NR (and less so model R) wrongly predicts a source in the high frequency group when a low frequency one is observed. The opposite is rarely the case, i.e. that a low frequency source is predicted when a high frequency source is observed. This is causing the average (Fig. 7c) of the matching observed PSDs to have more low frequency energy than the modeled ones. Including coastal reflections, as in model R shown in Fig. 8e, enhances the fit compared to model NR shown in Fig. 8d, because it adds low frequency energy to the source pressure-PSD. As we have seen from the average strongest-source PSDs



**Figure 9.** (a) Centroid beam-PSDs (black) #0–#4 that divide the dataset of matching normalized sources into five different groups. Red / blue / green are the modeled beam-PSDs averaged over the members of a particular cluster. They are ordered from lowest to highest dominant frequency. (b) Spatial distribution of clusters #0–#4. The observation limits of the arrays are indicated in (b) with faint black lines.

(Fig. 7b-c), the observed low frequency peak is also at slightly lower frequencies than the modeled one.

Modelled and observed strongest-source PSDs fit reasonably well in terms of the amplitudes and frequencies of their maxima. In the following, we examine the full shape of the strongest-source PSDs as a function of frequency and associate them with geographical regions.

### 3.4 Matching Strongest-Sources: Cluster Analysis

The existence of different groups of strongest-source PSDs with distinct spectral shape suggests that they can potentially be classified in a meaningful way. We can then examine the average spectral shape and geographical distribution of the individual classes instead of averaging over the entire dataset. To this end, we use a k-means cluster algorithm [Arthur and Vassilvitskii, 2007; Pedregosa *et al.*, 2011] that extracts representative PSDs from the observations, fully independent from the models. To focus on spectral shape rather than amplitude of the strongest-sources, we normalize strongest-source PSDs to have unit maximum. We then set the k-means algorithm to compute five mean spectra (called ‘centroids’) that minimize the summed L2 distances between the normalized observed strongest-source PSDs and the closest centroid.

Each day can then be labelled with the index #0-#4 of the centroid that the observed strongest-source PSD of this day is closest to (Table S1 in the supporting information material contains all labels). Importantly, the modeled spectra are labelled according to the centroid that the *observed* strongest-source PSD is closest to, as well. Our dataset of matching days is therefore subdivided into the five centroids based purely on the observations. We call the set of observed and modeled beam-PSDs that belong to the same centroid a ‘cluster’.

Fig. 9a shows the mean observed (black solid) and modeled (red/blue/green dotted) strongest-source PSDs of each cluster. For each cluster, the mean is taken over the subset

of days, that belong to the same cluster. Because the clusters were computed based on the observations, the mean observed beam-PSD of a cluster is the same as its centroid.

The clusters are ordered by the frequency of the maximum of their centroid PSD, i.e. by the observed strongest-source frequency, from lowest frequency (#0) to highest frequency (#4), as can be seen in Fig. 9a. The centroid PSDs #0 and #1 have a low frequency peak at 0.13Hz (7.7s) and 0.15Hz (6.7s), respectively. Centroid PSD #2 has a broad intermediate peak between 0.16Hz (6.7s) and 0.18Hz (5.5s), and the centroid PSDs #3 and #4 have a high frequency peak at 0.18Hz (5.5s) and 0.21Hz (4.8s), respectively. Clusters #0 and #1 can therefore be seen as low frequency dominated sources, and #3 and #4 as high frequency ones. Cluster #2 covers the intermediate frequency range.

Fig. 9b shows maps of the spatial distribution of these clusters from #0–#4, i.e. histograms of the number of observations that correspond to each cluster. The number of observations is biased by the coverage of the globe by the five arrays but also by the quality of the arrays that were used in this article. The spatial distribution of the clusters, allows the identification of spectral characteristics of particular regions.

The source region at the Greenland coast contains mostly low frequency sources of type #0–#2, predominantly of type #1. Other regions with mostly low frequency sources of type #1 are in the South-Central Pacific over the East-Pacific Rise. The low-frequency South-Central Pacific sources are particularly interesting because the centroid dominant frequency (0.15Hz) of cluster #1 corresponds to typical ridge depth resonance frequencies (see Fig. 3).

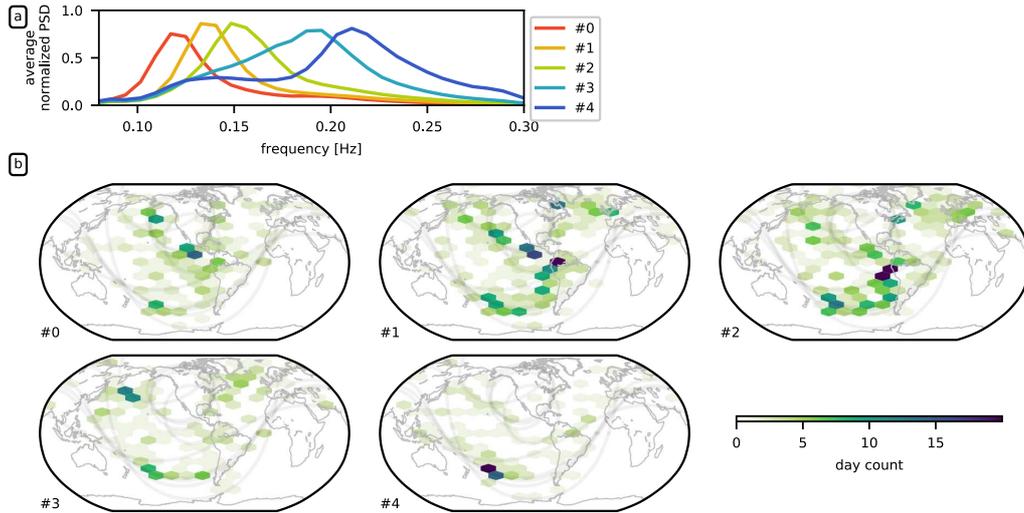
The low frequency centroids are not modeled satisfactory, which is what we already observed in Fig. 7 and Fig. 8. All modeled low frequency PSDs from clusters #0–#1 have significant high frequency energy around 0.20Hz, contrasting with the observed ones. The low frequency peaks of the observed centroid PSDs in #0 and #1 have slightly lower frequency than the modelled ones. These differences are smaller for model R that fits the centroid PSDs #0 and #1 better than model NR.

High frequency source #3 can mostly be found in the North-East Pacific along and north of the Japanese coast. This source region can be associated with typhoons that move from south to north along the coast. They emit P-waves with a characteristic high frequency dominated spectrum. Source type #4 can also be found in this region but also in the western South Pacific.

In contrast to the low frequency ones, high frequency sources are modeled quite accurately, in particular for cluster #3. The observed high frequency peak around 0.20Hz due to the source-site resonance frequency is modeled very accurately. The peak width of centroid PSD #3 is less wide than the peak width of centroid PSDs #0 and #1. This behaviour is reproduced in the models and is a consequence of the varying width of the resonance peaks (see Fig. 3a). The spectral peak of the centroid PSD #4 is not as narrow as the one of centroid PSD #3, possibly because these sources are not as homogeneous in spectral shape as cluster #3 sources. Also in the high frequency clusters, observed low frequency energy is stronger than the modeled one.

### 3.5 Non-Matching Strongest-Sources: Cluster Analysis

Finally, we also subdivide the non-matching observed sources (using the same approach as for the matching sources) to find five centroid spectra that minimize the L2 distance to the full dataset of non-matching normalized spectra. In this case, we cannot compare the observed and the modeled spectra because their slowness is associated with a different source region. Note that a notable part of these sources might correspond to other signals than secondary microseism P-waves, as we have explained previously. Therefore they have to be interpreted with care.



**Figure 10.** (a) Centroid beam-PSDs #0–#4 that divide the dataset of non-matching normalized strongest-sources into five different groups. They are ordered from lowest to highest dominant frequency. (b) Spatial distribution of clusters #0–#4. The observation limits of the arrays are indicated in (b) with faint black lines.

The cluster centroids shown in Fig. 10a are again ordered from lowest to highest dominant frequency. They split into a group of low frequency (#0–#2 from 0.12–0.16Hz) and high frequency (#3–#4 from 0.19–0.22Hz) clusters, again confirming that there are only few spectra with dominant frequency around 0.17Hz.

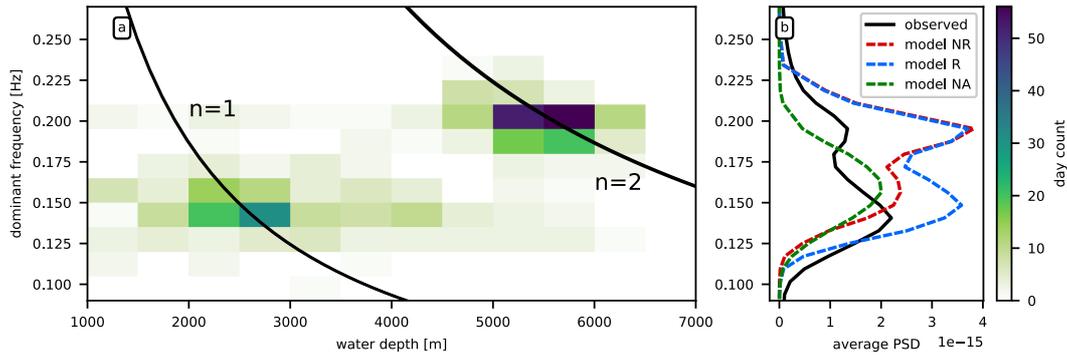
The spatial distribution of the clusters (Fig. 10b) shows a large group of low frequency sources along the Central and South American Pacific coast (#0–#2), with a particular high concentration at the North-Western tip of South America. Another accumulation of non-matching low frequency sources can be seen in the Central Pacific. As we have argued before, a large part of these sources cluster at the P-wave observation limits of the Mexico and Nevada arrays that suffer from high leakage in slowness domain. In our experience, a large part of these phases correspond to regional P-waves from sources closer than  $30^\circ$  that leak into the P-wave slowness range. They might therefore correspond to secondary microseisms, but are not mapped accurately and are not well modelled by ray theory.

Many low and high frequency sources are also located in the South Pacific. Here, the low frequency sources are spread from East to West over the South Pacific but the high frequency sources are concentrated in the West. Again, these sources concentrate around the P-wave observation limits of the arrays and likely correspond to P-waves that become influenced by the Earth’s core from sources at larger distances than  $90^\circ$ .

## 4 Discussion

Our analysis of secondary microseismic P-waves confirms source regions that have been seen and studied by many authors:

The North Atlantic, around Greenland, is known to have strong microseismic activity over the whole year and has been studied using surface and body waves [e.g. *Stehly et al.*, 2006; *Kedar et al.*, 2008; *Landès et al.*, 2010; *Stutzmann et al.*, 2012; *Obrebski et al.*, 2013; *Sergeant et al.*, 2013; *Euler et al.*, 2014; *Liu et al.*, 2016; *Nishida and Takagi*, 2016]. Microseismic sources from the North-West Pacific have also been studied in detail using different seismic phases and arrays [e.g. *Zhang et al.*, 2010a; *Landès et al.*, 2010; *Obrebski et al.*,



**Figure 11.** (a) Comparison of observed dominant frequency and water depth at the observed geographical origin of the matching strongest-sources. The resonance frequencies for slownesses in the  $30^\circ - 90^\circ$  distance range (takeoff angles of  $3.5^\circ - 7.9^\circ$ ) are shown with black lines. (b) shows the average PSDs of observations and models. Comparing with (a), it becomes evident that the two peaks are a combined effect of the wave energy distribution (see model NA in (b)) and the water column resonance frequencies.

2013; Pyle *et al.*, 2015; Gal *et al.*, 2015; Farra *et al.*, 2016], and – although the southern hemisphere is less well studied – a number of studies identified sources in the South Pacific, east of Australia [Landès *et al.*, 2010; Reading *et al.*, 2014; Euler *et al.*, 2014; Gal *et al.*, 2015].

To our knowledge, there are no studies of the coastal sources that we have observed along the Pacific coast of Central and South America. Further examination of these sources would be particularly interesting, because they seem to be most strongly related to coastal reflections (compare Fig. 5d with Fig. 5c), and therefore allow to isolate this phenomenon.

Contrary to the abundance of studies on source location, few authors examined the spectral characteristics of secondary microseismic sources. Many studies average over frequency bands that are too large to see the spectral source characteristics that we have observed. However, a general dependence of source location on frequency is a common observation [e.g. Zhang *et al.*, 2009; Euler *et al.*, 2014; Gal *et al.*, 2015].

Cluster analysis, as shown in this article, allows to separate observed sources with different spectra automatically and to associate them with different regions. Our two datasets, one with matching modeled sources and one without matching modeled sources, are divided into characteristic low frequency and high frequency clusters.

Several observations of dominant microseismic P-wave frequencies from North Pacific sources have been reported at 0.20Hz [Zhang *et al.*, 2010a; Obrebski *et al.*, 2013; Neale *et al.*, 2017; Farra *et al.*, 2016]. In our matching strongest-sources dataset these corresponds to cluster #3 sources that are located in the deep North Pacific as well. A strong P-wave source around Greenland has been seen with dominant frequency of approximately 0.13Hz [Nishida and Takagi, 2016], corresponding to cluster #0 that we have also found to be concentrated in the same region.

These independent observations of similar dominant frequencies in similar regions support our conclusion that the dominant frequency of secondary microseismic P-waves is to a large extent controlled by the source region. As we have shown, this can be explained and quantitatively modeled as the consequence of source-site resonance as formulated by [Arduin and Herbers, 2013; Gualtieri *et al.*, 2014; Farra *et al.*, 2016].

Importantly, source-site resonance, and therefore water depth, seems to have a stronger influence on the dominant frequency than the variations of the pressure source spectrum in-

duced by wave-wave interactions. Fig. 11a shows the relation between matching strongest-source frequency and the ocean height at their geographical position. We have used a bathymetry map smoothed over a spherical cap with radius of  $\sim 5$  degrees to extract the ocean height. Due to the limited resolution of the strongest-source position this is only an approximate depth.

As explained by Eq. 6, the shallow water peak corresponds to the first harmonic and the deep water peak to the second harmonic of P-wave resonance in the water column, shown in Fig. 11a with black lines. The average PSDs of the matching strongest-sources are shown in Fig. 11b. The PSD double peak in the observations as well as in the models NR and R lies at frequencies where the pressure-PSD source, exemplified by model NA, is amplified by resonance.

Although several articles have modeled secondary microseism surface wave spectra at isolated stations [e.g. *Kedar et al.*, 2008; *Ardhuin et al.*, 2011; *Stutzmann et al.*, 2012], synthetic secondary microseism P-wave spectra are yet uncommon. [*Neale et al.*, 2017] computed synthetic P-wave beam energy using an empirical regression formalism, whereas [*Farra et al.*, 2016; *Nishida and Takagi*, 2016] used wave propagation theory to compute synthetic beams, as we did in this article.

The results of this article show, that at least for the matching strongest-sources, modeled (models NR and R) and observed beam-PSDs are quite similar. Model R predicts the spectral shape better, in particular because it adds low frequency energy to the P-wave spectra compared to model NR. Also the general strongest-source distribution of model R is more similar to the observed one, in particular along the Pacific coast of South and Central America. However, model R cannot reconcile all the differences of model NR with observations. Model NA does not correctly predict the observed spectral characteristics. The remaining differences between model R and the observations, can be attributed to four principal causes: (i) the ocean wave model could be inaccurate. (ii) our model of Earth's structure, in particular under the source (ocean bottom) and under the array could be inaccurate. (iii) the plane wave assumption and ray theory that are used to compute the synthetic beam-PSD have limitations. (iv) Unmodeled signals add noise to the observations.

Apart from inaccuracies in the ocean wave model, that are difficult to estimate here, we expect heterogeneous Earth structure to be responsible for the largest part of the discrepancies between observed and modeled beams: Wavefronts are distorted by seismic heterogeneities, which lowers their coherence and therefore the beam-PSD amplitude. P-wave traveltimes among stations of an array are expected to be smaller than 1s [*Zhang et al.*, 2010b], so that the observed beam amplitude should be reduced by less than 20%. Furthermore, we have used a two-layer model with basaltic material under the ocean bottom to compute the amplification factor. This assumption may not be valid everywhere. Assuming a 2 layer model with sediment material under the ocean bottom, can greatly reduce the resonance amplification to very low values, as e.g. from a factor of 3.7 to 2.2 for soft sediments (parameters extracted from model Crust 1.0 [*Laske et al.*, 2013]:  $\rho_{\text{solid}} = 2300 \text{ kg/m}^3$ ,  $\beta_{\text{solid}} = 2000 \text{ m/s}$ ,  $\alpha_{\text{solid}} = 3300 \text{ m/s}$ ). Additional resonances within the sediment layer are also likely, which requires taking into account a realistic multi-layer medium under the ocean. Such a model has the potential to reconcile part of the differences that we have observed, in particular the mismatch of the observed and modelled low frequency peaks. Sediments have been empirically associated with variations in microseism amplitude [*Liu et al.*, 2016; *Nishida and Takagi*, 2016], and a detailed treatment will be necessary in future studies to improve our knowledge of the modeling error.

Much information is contained in the non-matching sources that were not used for comparison with the models in this article. *Neale et al.* [2017] were able to find 90% of matching observations focussing on the North Pacific only and excluding noisy observations beforehand. We think the selection rate of 26% of all days that we were able to use is much lower because:

(i) Secondary maxima are not compared in our dataset. When multiple strong sources exist, the models can easily predict a different strongest-source region than observed. Even days with small mismatches between models and observation can therefore be classified as non-matching. Given the range of source amplitudes (see Fig. 6) this can easily occur. A better selection criterion than the strongest daily source has the potential to provide a much larger dataset, which could for example allow one to compare the non-matching coastal sources to the models. On the other hand, the big advantage of using the strongest-source only is that its beam-PSD is least perturbed by other sources.

(ii) Noisy observations are included in our dataset. The median criterion that we have used can only remove outliers *within* a day, but not a whole day that is noisy due to e.g. a very large earthquake or local noise. A typical example would be a local storm that generates surface waves that raise the noise level high enough to mask secondary microseism P-waves at any slowness. On such a day, the strongest-source does not correspond to a secondary microseism but to a random fluctuation of leaked surface wave energy and will very likely end up in the non-matching dataset.

(iii) We locate observed sources from the maximum of the narrow-band smoothed PSD (0.04Hz). We were able to increase our dataset to 35% of all days using the broad-band smoothed PSD (width of about 0.1Hz). A plausible explanation for this is that uncorrelated noise is reduced by averaging over a larger band, whereas the correlated signal is retained. A broad-band only match is therefore an indicator for a noisy observation, which is why we have decided against using them.

## 5 Conclusions

We have characterized spectra (beam power spectral densities) of P-waves that were generated by secondary microseisms. We compiled a database of spectra that we analyzed in terms of amplitude, dominant frequency and geographical origin by systematic extraction of the strongest daily microseismic source from five years of vertical component data recorded by five North American arrays. We compared our observed spectra with synthetic datasets that were computed using an ocean wave hindcast [Ardhuin *et al.*, 2011] and the formalism developed by Farra *et al.* [2016].

The dominant frequencies of the observed P-wave spectra are concentrated in two distinct frequency bands, either around 0.13–0.15Hz or 0.19–0.21Hz, forming two spectral peaks. The sources can be classified by spectral shape using cluster analysis: Low frequency clusters (0.13–0.15Hz) are predominantly detected in the North Atlantic, in the north-central and in the southern Pacific, as well as along the South and Central American Pacific coast. High frequency clusters (0.19–0.21Hz) are mostly detected in deep ocean areas as in the North-Western Pacific and the southern Pacific away from the East Pacific rise.

Modelled sources that match with an observed source in time and location can be found mostly in the North Atlantic, as well as in the North-Western and South Pacific. Observed sources in the Central Pacific and along the Central and South American Pacific coast do not have matching strongest-source counterparts in the model. These sources may exist in the model but not as strongest-source of the day.

Amplitudes of the matching strongest-sources are predicted within a factor of 0.4-6 by the models. Dominant frequencies are also accurately predicted and turn out to be strongly influenced by the resonance frequency of the source-site water column. Water column resonance at ocean depths of about 2000-3000m (shallow ocean) corresponds to the first harmonic with dominant frequencies of about 0.13–0.15Hz, whereas resonance at ocean depths of about 5000-6000m (deep ocean) corresponds to the second harmonic and leads to dominant frequencies of about 0.20Hz.

The amplitude of secondary microseism P-waves depends on the combined strength of ocean wave interactions and source-site resonance. Both determine the principal source regions and spectral characteristics.

The accuracy of quantitative modeling of secondary microseisms is now sufficiently high to allow using standard inversion techniques to extract information about the secondary microseism sources and the subsurface.

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The following open scientific software packages were used: obspy [Krischer *et al.*, 2015], numpy & scipy [van der Walt *et al.*, 2011], scikitlearn [Pedregosa *et al.*, 2011] and matplotlib [Hunter *et al.*, 2007]. Numerical computations were performed on the S-CAPAD platform, IGP, France.

The facilities of IRIS Data Services, and specifically the IRIS Data Management Center, were used in this study for access to waveforms and related metadata from the following seismic networks: AK [Alaska Earthquake Center and Univ. of Alaska Fairbanks, 1987], TA [IRIS Transportable Array, 2003], AT [NOAA National Oceanic and Atmospheric Administration (USA), 1967], CI [California Institute of Technology (Caltech), 1926], ZA [West, 2006], XF [Grand and Ni, 2006], XC [James and Fouch, 2006], YX [Klemperer and Miller, 2010]. IRIS Data Services are funded through the Seismological Facilities for the Advancement of Geoscience and EarthScope (SAGE) Proposal of the National Science Foundation under Cooperative Agreement EAR-1261681.

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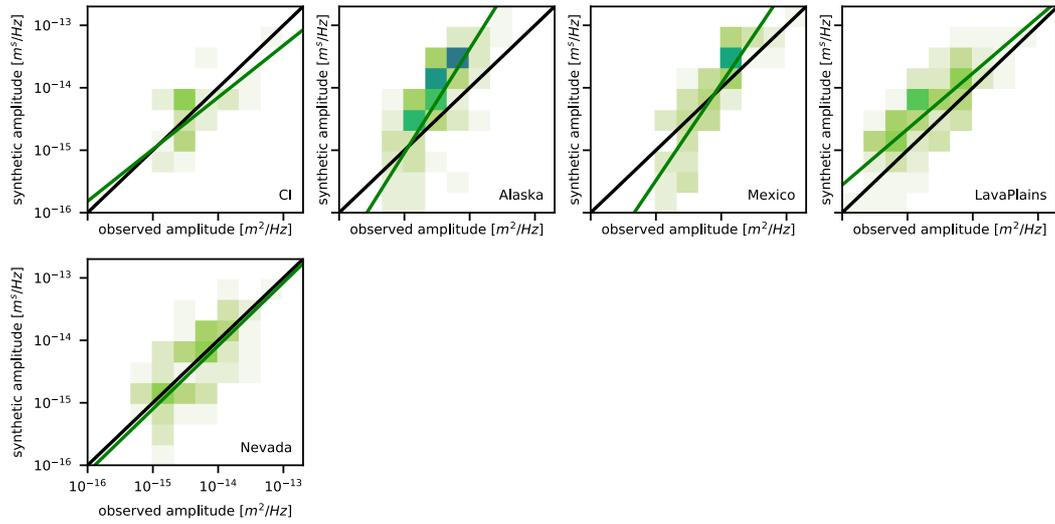
MS acknowledges MISTERIOS (CGL2013-48601-C2-1-R).

## A: Scatter plots for individual arrays

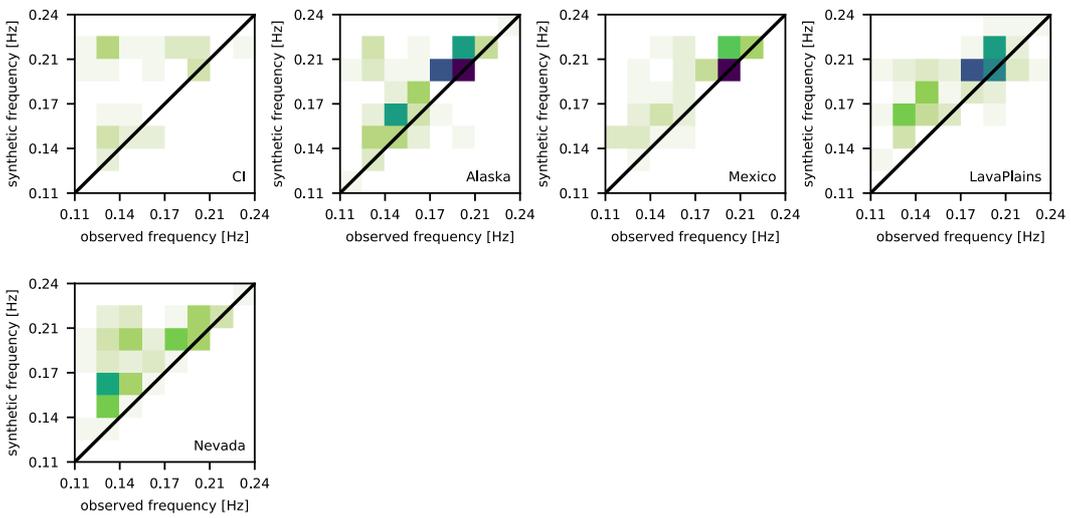
This appendix shows a comparison of the matching observed and synthetic (model R) strongest-source amplitudes A1 and frequencies A2 for each of the five arrays individually. Although each array observes slightly different sources, the overall observation of a spectral double peak, as well as the approximately valid amplitudes in the modeling process can be seen at each array.

## References

- Alaska Earthquake Center and Univ. of Alaska Fairbanks (1987), Alaska regional network, Seismic Network, doi:10.7914/SN/AK.
- Ardhuin, F., and T. H. Herbers (2013), Noise generation in the solid earth, oceans and atmosphere, from nonlinear interacting surface gravity waves in finite depth, *Journal of Fluid Mechanics*, 716, 316–348.
- Ardhuin, F., E. Stutzmann, M. Schimmel, and A. Mangeney (2011), Ocean wave sources of seismic noise, *Journal of Geophysical Research: Oceans*, 116(C9).
- Arthur, D., and S. Vassilvitskii (2007), k-means++: The advantages of careful seeding, in *Proceedings of the eighteenth annual ACM-SIAM symposium on Discrete algorithms*, pp. 1027–1035, Society for Industrial and Applied Mathematics.
- Bertelli, T. (1872), *Osservazioni sui piccoli movimenti dei pendoli in relazione ad alcuni fenomeni meteorologici del pd Timoteo Bertelli barnabita*, Tip. delle scienze matematiche



**Figure A1.** Comparison of the matching observed strongest-source amplitudes with model R for arrays CI (a), Alaska (b), Mexico (c), LavaPlains (d) and Nevada (e).



**Figure A2.** Comparison of the matching observed strongest-source frequencies with model R for arrays CI (a), Alaska (b), Mexico (c), LavaPlains (d) and Nevada (e).

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- California Institute of Technology (Caltech) (1926), Southern california seismic network, Other/Seismic Network, doi:10.7914/SN/CI.
- Ebeling, C. W. (2012), Inferring ocean storm characteristics from ambient seismic noise: A historical perspective, *Advances in Geophysics*, 53.
- Euler, G. G., D. A. Wiens, and A. A. Nyblade (2014), Evidence for bathymetric control on the distribution of body wave microseism sources from temporary seismic arrays in africa, *Geophysical Journal International*, p. ggu105.
- Farra, V., E. Stutzmann, L. Gualtieri, M. Schimmel, and F. Arduin (2016), Ray-theoretical modeling of secondary microseism p waves, *Geophysical Journal International*, 206(3), 1730–1739.
- Friedrich, A., F. Krüger, and K. Klinge (1998), Ocean-generated microseismic noise located with the gräfenberg array, *Journal of Seismology*, 2(1), 47–64.
- Gal, M., A. Reading, S. Ellingsen, L. Gualtieri, K. D. Koper, R. Burlacu, H. Tkalčić, and M. Hemer (2015), The frequency dependence and locations of short-period microseisms generated in the southern ocean and west pacific, *Journal of Geophysical Research: Solid Earth*, 120(8), 5764–5781.
- Gerstoft, P., K. G. Sabra, P. Roux, W. Kuperman, and M. C. Fehler (2006), Green’s functions extraction and surface-wave tomography from microseisms in southern california, *Geophysics*, 71(4), SI23–SI31.
- Gerstoft, P., P. M. Shearer, N. Harmon, and J. Zhang (2008), Global p, pp, and pkp wave microseisms observed from distant storms, *Geophysical Research Letters*, 35(23).
- Grand, S., and J. Ni (2006), Mapping the rivera subduction zone, Other/Seismic Network, doi:10.7914/SN/XF\_2006.
- Gualtieri, L., E. Stutzmann, V. Farra, Y. Capdeville, M. Schimmel, F. Arduin, and A. Morelli (2014), Modelling the ocean site effect on seismic noise body waves, *Geophysical Journal International*, p. ggu042.
- Hadziioannou, C., P. Gaebler, U. Schreiber, J. Wassermann, and H. Igel (2012), Examining ambient noise using colocated measurements of rotational and translational motion, *Journal of seismology*, 16(4), 787–796.
- Hasselmann, K. (1963), A statistical analysis of the generation of microseisms, *Reviews of Geophysics*, 1(2), 177–210.
- Haubrich, R. A., and K. McCamy (1969), Microseisms: Coastal and pelagic sources, *Reviews of Geophysics*, 7(3), 539–571.
- Hunter, J. D., et al. (2007), Matplotlib: A 2d graphics environment, *Computing in science and engineering*, 9(3), 90–95.
- IRIS Transportable Array (2003), Usarray transportable array, Other/Seismic Network, doi: 10.7914/SN/TA.
- James, D., and M. Fouch (2006), Collaborative research: Understanding the causes of continental intraplate tectonomagmatism: A case study in the pacific northwest, Other/Seismic Network, doi:10.7914/SN/XC\_2006.
- Kedar, S., M. Longuet-Higgins, F. Webb, N. Graham, R. Clayton, and C. Jones (2008), The origin of deep ocean microseisms in the north atlantic ocean, in *Proceedings of the Royal Society of London A: Mathematical, Physical and Engineering Sciences*, vol. 464, pp. 777–793, The Royal Society.
- Kennett, B., and E. Engdahl (1991), Traveltimes for global earthquake location and phase identification, *Geophysical Journal International*, 105(2), 429–465.
- Klemperer, S., and K. Miller (2010), Collaborative research: 4d multi-disciplinary investigation of highly variable crustal response to continental extension in the north-central basin and range, Other/Seismic Network, doi:10.7914/SN/YX\_2010.
- Koper, K. D., B. De Foy, and H. Benz (2009), Composition and variation of noise recorded at the yellowknife seismic array, 1991–2007, *Journal of Geophysical Research: Solid Earth*, 114(B10).

- Koper, K. D., K. Seats, and H. Benz (2010), On the composition of earth's short-period seismic noise field, *Bulletin of the Seismological Society of America*, *100*(2), 606–617, doi: 10.1785/0120090120.
- Krischer, L., T. Megies, R. Barsch, M. Beyreuther, T. Lecocq, C. Caudron, and J. Wassermann (2015), ObsPy: a bridge for seismology into the scientific Python ecosystem, *Computational Science & Discovery*, *8*(1), 14,003.
- Landès, M., F. Hubans, N. M. Shapiro, A. Paul, and M. Campillo (2010), Origin of deep ocean microseisms by using teleseismic body waves, *Journal of Geophysical Research: Solid Earth*, *115*(B5).
- Liu, Q., K. D. Koper, R. Burlacu, S. Ni, F. Wang, C. Zou, Y. Wei, M. Gal, and A. M. Reading (2016), Source locations of teleseismic P, SV, and SH waves observed in microseisms recorded by a large aperture seismic array in China, *Earth and Planetary Science Letters*, *449*, 39–47.
- Longuet-Higgins, M. S. (1950), A theory of the origin of microseisms, *Philosophical Transactions of the Royal Society of London A: Mathematical, Physical and Engineering Sciences*, *243*(857), 1–35.
- Meschede, M., and E. Stutzmann (2016), colormap2d-v0.5, Other/Python Package, doi: 10.5281/zenodo.192611.
- Montagner, J.-P., and B. Kennett (1996), How to reconcile body-wave and normal-mode reference earth models, *Geophysical Journal International*, *125*(1), 229–248.
- Neale, J., N. Harmon, and M. Srokosz (2017), Monitoring remote ocean waves using p-wave microseisms, *Journal of Geophysical Research: Oceans*.
- Nishida, K., and R. Takagi (2016), Teleseismic s wave microseisms, *Science*, *353*(6302), 919–921.
- Nishida, K., H. Kawakatsu, Y. Fukao, and K. Obara (2008), Background love and rayleigh waves simultaneously generated at the pacific ocean floors, *Geophysical Research Letters*, *35*(16).
- NOAA National Oceanic and Atmospheric Administration (USA) (1967), National tsunami warning center alaska seismic network, Other/Seismic Network, doi:10.7914/SN/AT.
- Obrebski, M., F. Arduin, E. Stutzmann, and M. Schimmel (2013), Detection of microseismic compressional (p) body waves aided by numerical modeling of oceanic noise sources, *Journal of Geophysical Research: Solid Earth*, *118*(8), 4312–4324.
- Pedregosa, F., G. Varoquaux, A. Gramfort, V. Michel, B. Thirion, O. Grisel, M. Blondel, P. Prettenhofer, R. Weiss, V. Dubourg, et al. (2011), Scikit-learn: Machine learning in python, *Journal of Machine Learning Research*, *12*(Oct), 2825–2830.
- Pyle, M. L., K. D. Koper, G. G. Euler, and R. Burlacu (2015), Location of high-frequency P wave microseismic noise in the Pacific Ocean using multiple small aperture arrays, *Geophysical Research Letters*, *42*(8), 2700–2708, doi:10.1002/2015GL063530.
- Rasche, N., and F. Arduin (2013), A global wave parameter database for geophysical applications. part 2: Model validation with improved source term parameterization, *Ocean Modelling*, *70*, 174–188.
- Reading, A. M., K. D. Koper, M. Gal, L. S. Graham, H. Tkalčić, and M. A. Hemer (2014), Dominant seismic noise sources in the southern ocean and west pacific, 2000–2012, recorded at the warramunga seismic array, australia, *Geophysical Research Letters*, *41*(10), 3455–3463.
- Rost, S., and C. Thomas (2002), Array seismology: Methods and applications, *Reviews of geophysics*, *40*(3).
- Schimmel, M., and J. Gallart (2007), Frequency-dependent phase coherence for noise suppression in seismic array data, *Journal of Geophysical Research: Solid Earth*, *112*(B4).
- Schimmel, M., and H. Paulssen (1997), Noise reduction and detection of weak, coherent signals through phase-weighted stacks, *Geophysical Journal International*, *130*(2), 497–505.
- Schimmel, M., E. Stutzmann, F. Arduin, and J. Gallart (2011), Polarized earth's ambient microseismic noise, *Geochemistry, Geophysics, Geosystems*, *12*(7).

- Schulte-Pelkum, V., P. S. Earle, and F. L. Vernon (2004), Strong directivity of ocean-generated seismic noise, *Geochemistry, Geophysics, Geosystems*, 5(3).
- Sergeant, A., E. Stutzmann, A. Maggi, M. Schimmel, F. Ardhuin, and M. Obrebski (2013), Frequency-dependent noise sources in the north atlantic ocean, *Geochemistry, Geophysics, Geosystems*, 14(12), 5341–5353.
- Stehly, L., M. Campillo, and N. Shapiro (2006), A study of the seismic noise from its long-range correlation properties, *Journal of Geophysical Research: Solid Earth*, 111(B10).
- Stutzmann, E., M. Schimmel, G. Patau, and A. Maggi (2009), Global climate imprint on seismic noise, *Geochemistry, Geophysics, Geosystems*, 10(11).
- Stutzmann, E., F. Ardhuin, M. Schimmel, A. Mangeney, and G. Patau (2012), Modelling long-term seismic noise in various environments, *Geophysical Journal International*, 191(2), 707–722, doi:10.1111/j.1365-246X.2012.05638.x.
- Tanimoto, T., C. Hadziioannou, H. Igel, J. Wasserman, U. Schreiber, and A. Gebauer (2015), Estimate of rayleigh-to-love wave ratio in the secondary microseism by colocated ring laser and seismograph, *Geophysical Research Letters*, 42(8), 2650–2655.
- Toksöz, M. N., and R. T. Lacoss (1968), Microseisms: Mode structure and sources, *Science*, 159(3817), 872–873.
- Tolman, H. L., et al. (2009), User manual and system documentation of wavewatch iii tm version 3.14, *Technical note, MMAB Contribution*, 276.
- van der Walt, S., S. C. Colbert, and G. Varoquaux (2011), The numpy array: A structure for efficient numerical computation, *Computing in Science Engineering*, 13(2), 22–30, doi: 10.1109/MCSE.2011.37.
- Vinnik, L. (1973), Sources of microseismic p-waves, *pure and applied geophysics*, 103(1), 282–289.
- West, M. (2006), The colima deep seismic experiment, Other/Seismic Network, doi: 10.7914/SN/ZA\_2006.
- Zhang, J., P. Gerstoft, and P. M. Shearer (2009), High-frequency P-wave seismic noise driven by ocean winds, *Geophysical Research Letters*, 36(9), 1–5, doi:10.1029/2009GL037761.
- Zhang, J., P. Gerstoft, and P. D. Bromirski (2010a), Pelagic and coastal sources of p-wave microseisms: Generation under tropical cyclones, *Geophysical Research Letters*, 37(15).
- Zhang, J., P. Gerstoft, and P. M. Shearer (2010b), Resolving p-wave travel-time anomalies using seismic array observations of oceanic storms, *Earth and Planetary Science Letters*, 292(3), 419–427.
- Laske, G., G. Masters, Z. Ma and M. Pasyanos (2013), Update on CRUST1. 0—A 1-degree global model of Earth's crust, *Geophys. Res. Abstracts*, 15, Abstract EGU2013-2658, 2013.

## Supporting Information for “The Effect of Water-Column Resonance on the Spectra of Secondary Microseism P-waves”

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### 1 Table of Matching Strongest Sources

Table S1 lists characteristics of the observed strongest-sources whose geographical origin matches with model R. The table columns correspond to:

- (i) the date when the strongest-source was observed.
- (ii) the geographical position that corresponds to the strongest-source slowness, assuming that it is a P-wave.
- (iii) the cluster label that is assigned to the strongest-source, based on the shape of its beam power spectral density (PSD). The cluster centroids that correspond to a certain type can be seen in Fig. 9a.
- (iv) the frequency of the beam-PSD maximum in [Hz].
- (v) the beam-PSD maximum in [ $m^2/Hz$ ].
- (vii) the normalized PSD in the frequency range from 0.08-0.3Hz. Faint vertical ticks indicate the position of 0.10Hz, 0.15Hz, 0.20Hz, and 0.25Hz.

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**Table S1:** observed matching strongest-source properties

date	origin	type	freq.	max.	PSD
06-1-24	59.72N -39.35E	1	0.15	1e-14	
06-1-30	38.76N 178.12E	2	0.20	3e-15	
06-1-31	41.65N -178.27E	3	0.20	2e-15	
06-4-22	-43.47N -36.96E	3	0.19	2e-15	
06-4-23	-43.94N -43.26E	2	0.15	1e-15	
06-5-21	-55.80N -137.31E	1	0.16	2e-15	
06-5-22	-45.78N -137.55E	1	0.16	2e-15	
06-5-29	-40.85N -145.62E	2	0.17	2e-15	
06-6-25	-45.49N -130.11E	1	0.14	3e-15	
06-6-26	-52.25N -131.59E	1	0.15	2e-15	
06-6-27	-39.10N -38.54E	3	0.20	2e-15	
06-7-3	-39.58N -162.86E	0	0.13	2e-15	
06-7-4	-39.58N -162.86E	0	0.13	2e-15	
06-7-8	-4.44N -83.57E	0	0.14	3e-15	
06-7-10	-50.86N -57.36E	1	0.13	2e-15	
06-7-11	-52.33N -50.23E	1	0.14	2e-15	
06-7-18	-43.47N -170.24E	0	0.10	1e-15	
06-7-21	-50.86N -149.84E	1	0.13	1e-15	
06-7-22	-50.86N -149.84E	1	0.14	1e-15	
06-7-24	-43.47N -36.96E	2	0.20	7e-16	
06-7-25	-39.34N -156.99E	4	0.23	1e-15	
06-7-26	-48.62N -143.31E	0	0.13	5e-15	
06-7-27	-52.33N -140.87E	0	0.13	3e-14	
06-7-28	-52.33N -140.87E	0	0.14	1e-14	
06-7-29	-43.94N -43.26E	2	0.19	2e-15	
06-8-19	-34.30N -150.61E	4	0.20	8e-16	
06-8-20	-52.54N -118.11E	0	0.14	1e-15	
06-8-29	-56.70N -51.87E	1	0.14	3e-15	
06-9-3	-41.31N -116.94E	1	0.15	1e-15	
06-9-26	41.65N -178.27E	0	0.12	1e-14	
06-10-6	-43.63N -49.59E	2	0.17	2e-15	
06-10-9	37.24N -161.65E	4	0.20	8e-16	
06-10-12	40.17N -172.65E	4	0.20	7e-16	
06-10-22	-40.85N -145.62E	0	0.12	6e-15	
06-10-31	-41.49N -81.22E	1	0.15	2e-15	
06-11-27	53.50N -3.55E	2	0.18	2e-15	
06-12-21	55.88N -147.20E	1	0.15	9e-15	
08-1-4	62.14N -48.65E	2	0.17	7e-15	
08-1-7	46.22N -178.94E	3	0.20	5e-15	
08-1-8	43.88N 178.81E	3	0.20	4e-15	
08-1-11	45.33N 174.16E	3	0.19	4e-15	
08-1-12	41.45N 177.24E	3	0.19	6e-15	
08-1-14	54.37N 174.06E	2	0.16	2e-15	
08-1-15	60.04N -41.01E	2	0.17	1e-15	
08-1-18	41.45N 177.24E	3	0.20	4e-15	
08-1-19	43.88N 178.81E	3	0.20	4e-15	
08-1-20	41.45N 177.24E	3	0.20	3e-15	
08-1-21	43.88N 178.81E	3	0.20	5e-15	
08-1-23	42.71N 172.30E	3	0.20	3e-15	
08-1-24	45.33N 174.16E	3	0.20	4e-15	
08-1-25	45.33N 174.16E	3	0.20	3e-15	
08-1-26	48.66N 159.30E	2	0.18	5e-15	
08-1-28	66.01N -42.61E	1	0.16	5e-15	
08-1-29	74.75N -38.64E	1	0.15	2e-15	
08-1-31	64.50N -25.57E	0	0.14	6e-15	
08-2-2	62.14N -48.65E	1	0.14	3e-15	
08-2-3	62.14N -48.65E	1	0.16	4e-15	
08-2-4	38.57N 154.17E	2	0.19	3e-15	
08-2-5	44.33N 167.63E	2	0.14	2e-15	
08-2-8	57.38N -34.89E	1	0.14	2e-14	
08-2-9	62.14N -48.65E	0	0.14	7e-15	
08-2-12	41.45N 177.24E	3	0.19	4e-15	
08-2-14	44.33N 167.63E	3	0.20	4e-15	
08-2-17	49.26N 166.00E	3	0.19	2e-15	
08-2-18	49.26N 166.00E	3	0.20	8e-16	
08-2-23	41.41N 166.05E	2	0.19	1e-15	
08-2-24	36.15N -172.69E	3	0.20	3e-15	
08-2-28	51.33N 155.72E	1	0.14	2e-15	
08-2-29	49.26N 166.00E	3	0.20	2e-15	

Continued table

date	origin	type	freq.	max.	PSD
08-3-1	49.75N 172.88E	3	0.20	2e-15	
08-3-3	47.82N 176.71E	3	0.20	3e-15	
08-3-5	64.81N -50.26E	1	0.16	2e-15	
08-3-6	62.14N -48.65E	1	0.15	4e-15	
08-3-7	63.03N -41.04E	1	0.16	1e-15	
08-3-8	63.03N -41.04E	1	0.16	1e-15	
08-3-9	60.04N -41.01E	1	0.16	7e-16	
08-3-13	46.22N -178.94E	3	0.20	2e-15	
08-3-17	42.80N -61.36E	2	0.16	1e-15	
08-3-18	43.88N -56.29E	2	0.17	1e-15	
08-3-21	64.50N -25.57E	1	0.16	8e-16	
08-3-27	63.81N -33.34E	2	0.17	6e-16	
08-3-28	63.03N -41.04E	1	0.16	1e-15	
08-3-29	63.03N -41.04E	1	0.16	1e-15	
08-3-30	60.94N -26.05E	1	0.16	1e-15	
08-4-5	26.52N 165.74E	3	0.20	5e-16	
08-4-6	-38.08N -160.71E	4	0.20	4e-16	
08-4-8	35.62N 142.88E	2	0.19	6e-16	
08-4-9	33.33N 148.21E	3	0.19	2e-15	
08-4-10	35.05N 153.68E	3	0.20	5e-16	
08-4-12	45.42N 156.90E	3	0.20	8e-16	
08-4-13	51.69N 162.54E	3	0.20	3e-15	
08-4-16	41.45N 177.24E	3	0.20	8e-16	
08-4-20	48.40N -175.97E	3	0.20	8e-16	
08-4-21	48.40N -175.97E	3	0.20	4e-16	
08-4-26	5.63N -31.68E	0	0.12	3e-15	
08-4-27	5.63N -31.68E	0	0.13	2e-15	
08-4-28	5.63N -31.68E	1	0.14	4e-16	
08-5-12	30.08N 137.70E	2	0.16	1e-15	
08-5-13	31.65N 142.89E	3	0.19	1e-15	
08-5-17	-40.25N -110.48E	1	0.15	1e-15	
08-5-27	-38.77N -144.39E	4	0.21	4e-16	
08-5-28	-41.35N -139.75E	4	0.21	5e-16	
08-5-29	-40.90N -121.53E	1	0.15	7e-16	
08-5-31	-38.77N -144.39E	4	0.23	4e-16	
08-6-1	-40.90N -121.53E	1	0.12	2e-16	
08-6-2	31.65N 142.89E	3	0.19	1e-16	
08-6-3	35.62N 142.88E	3	0.19	4e-16	
08-6-4	33.33N 148.21E	3	0.19	6e-16	
08-6-7	-28.80N -83.58E	3	0.18	7e-16	
08-6-9	-33.81N -163.05E	4	0.20	7e-16	
08-6-10	-33.81N -163.05E	4	0.20	8e-16	
08-6-26	-39.05N -152.46E	4	0.20	3e-16	
08-6-30	-36.03N -121.14E	1	0.16	8e-16	
08-7-4	-38.77N -93.10E	2	0.16	3e-16	
08-7-6	28.86N -157.99E	3	0.20	2e-16	
08-7-11	-44.84N -109.08E	0	0.13	4e-15	
08-7-12	-45.58N -115.47E	1	0.14	1e-15	
08-7-13	-39.05N -152.46E	0	0.13	7e-16	
08-7-15	-38.77N -144.39E	4	0.20	8e-16	
08-7-23	-29.56N -164.75E	4	0.20	3e-16	
08-7-24	-27.53N -160.08E	4	0.21	3e-16	
08-7-25	-33.81N -163.05E	4	0.20	3e-16	
08-7-26	-31.57N -158.26E	4	0.21	6e-16	
08-7-29	-32.42N -151.44E	4	0.20	7e-16	
08-7-30	-35.77N -148.28E	4	0.20	5e-16	
08-8-2	-35.45N -111.63E	2	0.20	1e-16	
08-8-3	-36.03N -116.34E	2	0.19	1e-16	
08-8-4	-36.03N -116.34E	2	0.18	1e-16	
08-8-16	-35.45N -125.86E	1	0.15	4e-16	
08-8-20	-42.53N -148.38E	4	0.20	6e-16	
08-8-21	-43.41N -134.39E	2	0.14	7e-16	
08-8-23	-40.90N -115.95E	1	0.15	6e-16	
08-8-24	-40.90N -115.95E	1	0.16	2e-16	
08-9-24	63.03N -41.04E	2	0.16	6e-16	
08-9-25	46.32N 163.29E	3	0.19	7e-16	
08-10-8	59.44N -48.52E	0	0.13	4e-15	
08-10-9	60.04N -41.01E	1	0.15	1e-14	
08-10-10	63.03N -41.04E	1	0.16	2e-15	
08-10-20	61.27N -18.57E	1	0.16	2e-15	
08-10-21	43.23N 161.34E	3	0.19	1e-15	
08-10-22	47.13N 169.88E	3	0.19	2e-15	

Continued table						Continued table					
date	origin	type	freq.	max.	PSD	date	origin	type	freq.	max.	PSD
08-10-24	63.03N -41.04E	1	0.15	1e-15		11-4-19	40.29N -177.50E	4	0.20	1e-15	
08-10-27	74.06N -13.14E	1	0.14	2e-15		11-4-24	57.85N -21.25E	2	0.17	1e-15	
08-11-6	47.82N 176.71E	3	0.20	2e-15		11-4-25	57.85N -21.25E	2	0.16	6e-16	
08-11-7	47.82N 176.71E	3	0.20	1e-15		11-4-26	-39.97N -158.84E	4	0.22	7e-16	
08-11-8	45.42N 156.90E	3	0.20	6e-16		11-5-3	38.39N 159.49E	3	0.20	1e-15	
08-11-9	44.33N 167.63E	3	0.19	5e-15		11-5-12	40.71N 171.48E	4	0.20	1e-15	
08-11-10	45.33N 174.16E	3	0.20	4e-15		11-5-13	39.08N 176.28E	3	0.20	1e-15	
08-11-11	43.88N 178.81E	3	0.20	1e-15		11-5-14	36.35N 175.37E	4	0.20	9e-16	
08-11-19	42.80N -176.13E	3	0.20	6e-16		11-6-30	-48.14N -118.90E	2	0.16	1e-15	
08-11-20	42.80N -176.13E	3	0.20	2e-15		11-7-4	-44.74N -146.33E	2	0.15	1e-15	
08-11-21	51.97N 169.56E	2	0.16	2e-15		11-7-6	-28.12N -168.39E	4	0.20	1e-15	
08-11-23	54.37N 174.06E	2	0.17	1e-15		11-7-12	11.49N -98.10E	0	0.12	2e-15	
08-11-24	54.16N -178.29E	2	0.17	2e-15		11-7-14	-37.99N -122.82E	2	0.16	2e-15	
08-11-25	37.09N 148.46E	3	0.19	2e-15		11-7-15	-43.46N -118.37E	1	0.16	3e-15	
08-11-26	42.71N 172.30E	3	0.20	3e-15		11-8-1	-44.74N -146.33E	4	0.20	2e-15	
08-11-27	46.22N -178.94E	3	0.20	3e-15		11-8-5	-39.97N -158.84E	4	0.21	2e-15	
08-11-28	48.84N -167.53E	0	0.13	3e-15		11-8-8	-48.14N -112.03E	1	0.15	3e-15	
08-12-1	47.82N 176.71E	0	0.13	4e-15		11-8-9	-47.36N -105.34E	1	0.14	4e-15	
08-12-2	48.40N -175.97E	2	0.14	2e-15		11-8-14	-44.74N -146.33E	4	0.20	2e-15	
08-12-3	42.12N -170.49E	3	0.20	1e-15		11-8-15	-39.97N -158.84E	4	0.21	1e-15	
08-12-4	64.81N -50.26E	1	0.16	2e-15		11-8-16	-44.74N -146.33E	2	0.20	1e-15	
08-12-5	29.41N 164.97E	3	0.19	3e-15		11-9-12	50.92N -26.38E	1	0.15	3e-15	
08-12-6	31.55N 175.95E	4	0.20	1e-15		11-9-13	57.85N -21.25E	2	0.17	2e-15	
08-12-7	38.12N -172.45E	3	0.20	3e-15		11-10-4	69.67N -42.27E	1	0.15	2e-15	
08-12-8	42.12N -170.49E	3	0.20	1e-15		11-10-6	64.52N -27.68E	1	0.16	2e-15	
08-12-10	62.14N -48.65E	1	0.15	2e-15		11-10-15	10.84N -107.20E	0	0.13	2e-15	
08-12-12	33.62N -165.32E	3	0.20	3e-15		11-10-23	34.93N -161.50E	3	0.20	9e-16	
08-12-15	46.32N 163.29E	3	0.19	4e-15		11-10-30	56.48N -42.67E	1	0.14	4e-15	
08-12-18	63.03N -41.04E	0	0.12	8e-16		11-11-17	58.21N -155.89E	1	0.16	4e-15	
08-12-22	57.38N -34.89E	1	0.16	3e-15		11-11-28	66.70N -37.47E	1	0.15	5e-15	
08-12-23	59.44N -48.52E	0	0.13	7e-15		11-11-30	68.06N -29.90E	1	0.14	2e-15	
08-12-27	45.42N 156.90E	2	0.20	2e-15		11-12-1	63.26N 2.93E	2	0.16	3e-15	
08-12-28	43.23N 161.34E	2	0.19	2e-15		11-12-2	36.99N -173.34E	3	0.20	5e-15	
08-12-30	47.13N 169.88E	3	0.19	3e-15		11-12-3	36.99N -173.34E	3	0.20	5e-15	
08-12-31	47.13N 169.88E	3	0.19	3e-15		11-12-4	40.29N -177.50E	3	0.20	9e-15	
11-1-1	40.71N 171.48E	3	0.20	9e-15		11-12-5	41.74N 177.73E	3	0.20	7e-15	
11-1-2	43.55N 173.27E	3	0.20	1e-14		11-12-6	40.71N 171.48E	3	0.20	4e-15	
11-1-3	43.55N 173.27E	3	0.20	7e-15		11-12-8	58.47N -6.86E	2	0.16	2e-15	
11-1-4	40.71N 171.48E	3	0.20	4e-15		11-12-10	58.21N -75.04E	0	0.12	4e-14	
11-1-6	42.72N 166.96E	3	0.19	6e-15		11-12-14	42.65N -175.79E	3	0.20	8e-15	
11-1-8	39.21N -172.39E	3	0.20	7e-15		11-12-15	44.91N -173.41E	3	0.20	7e-15	
11-1-9	40.29N -177.50E	3	0.20	5e-15		11-12-23	44.91N -173.41E	4	0.20	6e-15	
11-1-10	42.65N -175.79E	3	0.20	8e-15		11-12-27	41.74N 177.73E	3	0.20	1e-14	
11-1-16	42.65N -175.79E	3	0.20	9e-15		11-12-28	39.08N 176.28E	3	0.20	6e-15	
11-1-17	44.29N 179.77E	3	0.19	9e-15		11-12-29	36.35N 175.37E	3	0.20	4e-15	
11-1-27	41.74N 177.73E	3	0.19	7e-15		15-1-1	61.62N -32.87E	2	0.16	2e-15	
11-1-28	44.91N -173.41E	3	0.20	5e-15		15-1-5	65.30N -36.08E	1	0.14	4e-15	
11-1-29	45.39N -166.01E	4	0.21	4e-15		15-1-5	61.62N -32.87E	1	0.15	4e-15	
11-1-30	42.65N -175.79E	4	0.20	4e-15		15-1-6	41.48N -163.82E	0	0.13	1e-14	
11-1-31	41.38N -170.92E	3	0.20	1e-14		15-1-7	61.28N -31.97E	1	0.15	6e-15	
11-2-1	43.46N -168.84E	3	0.20	1e-14		15-1-7	61.62N -32.87E	1	0.14	8e-15	
11-2-2	60.22N -34.71E	1	0.14	7e-15		15-1-8	61.62N -32.87E	1	0.15	4e-15	
11-2-6	36.99N -173.34E	3	0.20	8e-15		15-1-8	61.28N -31.97E	2	0.14	3e-15	
11-2-7	36.57N -167.61E	3	0.20	8e-15		15-1-9	59.88N -18.45E	0	0.14	2e-15	
11-2-8	38.54N -166.83E	3	0.20	6e-15		15-1-10	59.88N -18.45E	0	0.14	3e-15	
11-2-11	61.17N -49.78E	0	0.14	2e-14		15-1-10	58.25N -19.47E	0	0.12	1e-15	
11-2-12	59.40N -41.99E	0	0.14	5e-14		15-1-11	56.29N -12.94E	0	0.14	3e-15	
11-2-13	60.95N -27.39E	0	0.12	3e-14		15-1-11	56.83N -13.73E	0	0.12	1e-15	
11-2-15	59.40N -41.99E	0	0.12	1e-14		15-1-12	56.83N -13.73E	0	0.12	2e-15	
11-2-23	59.40N -41.99E	1	0.16	6e-15		15-1-12	59.88N -18.45E	1	0.14	2e-15	
11-2-24	59.40N -41.99E	0	0.14	3e-14		15-1-13	61.62N -32.87E	1	0.14	6e-15	
11-2-25	61.17N -49.78E	0	0.14	8e-15		15-1-14	55.44N -23.51E	2	0.16	4e-15	
11-2-26	41.74N 177.73E	3	0.20	6e-15		15-1-14	58.87N -25.63E	1	0.15	3e-15	
11-3-4	42.72N 166.96E	3	0.19	2e-15		15-1-15	58.87N -25.63E	1	0.14	5e-15	
11-3-5	46.25N 175.72E	3	0.20	1e-15		15-1-15	56.83N -13.73E	2	0.16	6e-15	
11-3-6	46.67N -177.53E	4	0.20	1e-15		15-1-16	58.87N -25.63E	1	0.14	6e-15	
11-3-7	46.96N -170.35E	4	0.20	3e-15		15-1-16	58.25N -19.47E	2	0.16	7e-15	
11-3-8	46.25N 175.72E	3	0.20	2e-15		15-1-18	57.81N -32.62E	1	0.16	3e-15	
11-3-28	40.71N 171.48E	3	0.20	6e-15		15-1-18	56.88N -39.65E	2	0.16	2e-15	
11-3-29	41.74N 177.73E	3	0.20	6e-15		15-1-19	58.87N -25.63E	1	0.14	5e-15	
11-4-15	40.71N 171.48E	3	0.19	2e-15		15-1-19	61.28N -31.97E	1	0.14	4e-15	
11-4-18	41.74N 177.73E	4	0.20	1e-15		15-1-20	58.87N -25.63E	1	0.14	3e-15	

Continued table						Continued table					
date	origin	type	freq.	max.	PSD	date	origin	type	freq.	max.	PSD
15-1-21	57.00N -5.82E	0	0.14	3e-15		15-5-31	62.65N -22.07E	4	0.20	1e-16	
15-1-23	39.35N 153.57E	3	0.19	3e-15		15-6-1	62.65N -22.07E	2	0.20	1e-16	
15-1-26	61.62N -32.87E	0	0.13	6e-15		15-6-12	-36.28N -149.04E	2	0.16	6e-16	
15-1-28	72.04N -0.76E	1	0.14	2e-15		15-6-19	-39.40N -153.61E	4	0.21	5e-16	
15-1-28	67.85N 6.47E	1	0.14	4e-15		15-6-20	-35.78N -142.32E	4	0.23	3e-16	
15-1-29	72.61N -8.24E	0	0.12	8e-15		15-6-24	-30.82N -167.93E	4	0.20	1e-15	
15-1-29	65.55N -10.42E	0	0.12	5e-15		15-6-25	-35.23N -167.46E	4	0.20	6e-16	
15-1-30	65.55N -10.42E	0	0.12	2e-15		15-6-26	-41.98N -158.74E	4	0.21	5e-16	
15-1-30	37.64N 167.93E	3	0.19	3e-15		15-6-29	-42.99N -84.30E	0	0.13	4e-16	
15-2-6	42.07N 164.57E	2	0.20	1e-15		15-6-30	-32.73N -83.66E	0	0.13	3e-16	
15-2-15	54.92N -33.56E	2	0.14	3e-15		15-7-2	14.25N 156.80E	4	0.21	3e-16	
15-2-17	57.00N -5.82E	1	0.14	1e-15		15-7-3	8.80N 154.19E	4	0.20	4e-16	
15-2-18	57.81N -32.62E	1	0.14	9e-15		15-7-3	14.01N 150.09E	4	0.19	3e-16	
15-2-18	58.91N -35.61E	1	0.14	9e-15		15-7-4	12.61N 152.43E	3	0.20	5e-16	
15-2-19	58.87N -25.63E	0	0.14	2e-14		15-7-4	19.10N 148.83E	4	0.20	3e-16	
15-2-19	61.28N -31.97E	0	0.14	2e-14		15-7-8	14.82N 146.42E	2	0.16	8e-16	
15-2-22	63.97N -28.79E	2	0.20	6e-15		15-7-10	27.35N 132.31E	3	0.20	5e-16	
15-2-22	58.87N -25.63E	1	0.16	4e-15		15-7-12	23.51N 137.24E	4	0.20	2e-16	
15-2-23	58.25N -19.47E	0	0.13	3e-14		15-7-15	27.85N 140.39E	4	0.20	2e-16	
15-2-23	58.87N -25.63E	0	0.14	3e-14		15-7-22	-21.16N -80.51E	0	0.12	7e-16	
15-2-24	58.87N -25.63E	1	0.15	4e-15		15-7-25	-42.58N -104.40E	0	0.14	2e-15	
15-2-24	61.38N -15.78E	1	0.15	3e-15		15-7-26	-24.84N -103.58E	0	0.14	1e-15	
15-2-25	58.87N -25.63E	1	0.16	2e-15		15-7-27	-15.75N -92.04E	1	0.16	1e-15	
15-2-25	59.74N -25.52E	2	0.16	1e-15		15-8-5	21.98N 139.96E	4	0.22	2e-16	
15-2-26	58.87N -25.63E	1	0.14	8e-15		15-8-6	21.04N 131.50E	3	0.20	6e-16	
15-2-26	61.28N -31.97E	1	0.14	1e-14		15-8-6	-46.28N -88.23E	0	0.13	5e-16	
15-2-27	57.81N -32.62E	0	0.13	1e-14		15-8-7	23.22N 128.91E	3	0.20	5e-16	
15-2-27	58.91N -35.61E	0	0.13	1e-14		15-8-8	25.63N 126.53E	3	0.20	2e-16	
15-2-28	58.87N -25.63E	1	0.14	3e-15		15-8-12	-39.58N -146.15E	4	0.20	7e-16	
15-2-28	59.74N -25.52E	1	0.16	1e-15		15-8-13	61.28N -31.97E	1	0.16	4e-16	
15-3-1	58.87N -25.63E	2	0.17	1e-15		15-8-14	-39.58N -146.15E	4	0.21	6e-16	
15-3-2	58.87N -25.63E	2	0.16	2e-15		15-8-16	19.56N 161.44E	4	0.20	2e-16	
15-3-3	58.87N -25.63E	2	0.16	9e-16		15-8-17	14.70N 156.98E	3	0.20	1e-15	
15-3-4	51.58N -42.72E	1	0.14	2e-15		15-8-17	19.56N 161.44E	3	0.20	3e-16	
15-3-4	49.92N -42.10E	1	0.15	3e-15		15-8-19	12.61N 152.43E	3	0.18	3e-15	
15-3-5	54.92N -33.56E	1	0.16	1e-15		15-8-19	23.42N 154.05E	3	0.20	7e-16	
15-3-5	56.70N -39.44E	0	0.12	2e-15		15-8-20	20.50N 149.99E	3	0.20	1e-15	
15-3-6	62.84N -38.94E	1	0.16	2e-15		15-8-20	24.36N 148.28E	3	0.19	5e-16	
15-3-6	57.81N -32.62E	1	0.14	2e-15		15-8-21	25.25N 145.53E	3	0.19	8e-16	
15-3-7	61.28N -31.97E	1	0.16	2e-15		15-8-21	23.28N 144.31E	3	0.19	1e-15	
15-3-7	58.87N -25.63E	1	0.14	3e-15		15-8-22	26.42N 142.89E	4	0.20	7e-16	
15-3-8	63.97N -28.79E	1	0.16	2e-15		15-8-23	-49.34N -120.69E	1	0.14	3e-15	
15-3-8	61.62N -32.87E	1	0.14	2e-15		15-8-23	25.25N 145.53E	4	0.23	4e-15	
15-3-9	58.87N -25.63E	1	0.14	1e-15		15-8-24	32.50N 149.46E	3	0.19	1e-15	
15-3-10	40.49N 151.72E	3	0.18	1e-15		15-8-24	35.09N 149.57E	4	0.18	9e-16	
15-3-10	35.84N 144.55E	3	0.19	2e-15		15-8-25	38.44N 155.60E	3	0.19	1e-15	
15-3-12	58.87N -25.63E	1	0.16	3e-15		15-8-25	32.50N 149.46E	3	0.19	1e-15	
15-3-12	58.25N -19.47E	1	0.16	3e-15		15-8-26	42.93N 160.26E	3	0.20	8e-16	
15-3-14	58.87N -25.63E	1	0.14	3e-15		15-8-26	32.50N 149.46E	1	0.16	1e-15	
15-3-15	59.88N -18.45E	1	0.14	3e-15		15-8-31	23.29N -178.54E	4	0.20	4e-16	
15-3-16	61.62N -32.87E	1	0.16	2e-15		15-9-1	-39.47N -119.76E	1	0.15	1e-14	
15-3-19	61.62N -32.87E	1	0.15	6e-15		15-9-2	12.98N -143.64E	4	0.20	6e-16	
15-3-19	61.28N -31.97E	1	0.15	4e-15		15-9-3	12.98N -143.64E	4	0.20	9e-16	
15-3-20	61.62N -32.87E	1	0.14	5e-15		15-9-4	12.98N -143.64E	4	0.20	5e-16	
15-3-20	61.28N -31.97E	1	0.16	2e-15		15-9-6	27.57N 172.14E	4	0.19	5e-16	
15-3-21	54.92N -33.56E	2	0.15	9e-16		15-9-7	30.78N 170.97E	3	0.19	6e-16	
15-3-21	57.81N -32.62E	1	0.15	2e-15		15-9-8	32.86N 164.67E	3	0.19	1e-15	
15-3-22	-49.34N -114.21E	1	0.16	1e-15		15-9-11	36.41N 150.30E	3	0.19	2e-15	
15-3-22	40.39N -27.13E	2	0.16	3e-16		15-9-14	36.58N -179.32E	4	0.20	5e-16	
15-3-25	-49.70N -83.16E	1	0.16	3e-15		15-9-14	37.34N -175.38E	4	0.20	8e-16	
15-3-31	60.19N -9.83E	2	0.16	9e-16		15-9-15	36.58N -179.32E	4	0.21	2e-16	
15-4-13	57.15N -29.36E	2	0.16	2e-15		15-9-19	54.92N -33.56E	4	0.20	9e-17	
15-4-21	41.55N -37.19E	2	0.15	1e-15		15-9-20	58.87N -25.63E	2	0.17	3e-16	
15-4-22	45.36N -36.92E	4	0.14	5e-16		15-9-21	42.36N 162.30E	4	0.20	2e-16	
15-4-25	29.95N -167.50E	4	0.20	5e-16		15-9-21	37.03N 156.15E	3	0.20	2e-16	
15-4-29	44.30N -42.10E	4	0.21	4e-16		15-9-22	42.93N 160.26E	3	0.20	3e-16	
15-5-11	59.88N -18.45E	2	0.16	1e-15		15-9-23	42.93N 160.26E	4	0.20	4e-16	
15-5-11	56.83N -13.73E	2	0.18	6e-16		15-9-23	40.57N 157.81E	3	0.20	6e-16	
15-5-12	60.19N -9.83E	2	0.18	4e-16		15-9-25	43.38N 173.06E	3	0.20	3e-16	
15-5-16	63.97N -28.79E	2	0.18	3e-16		15-9-26	25.30N 134.68E	3	0.20	2e-16	
15-5-17	59.74N -25.52E	2	0.18	7e-16		15-9-27	25.30N 134.68E	4	0.19	1e-16	
15-5-18	-45.89N -155.72E	1	0.14	2e-15		15-10-4	20.43N 164.59E	4	0.20	9e-16	

Continued table

date	origin	type	freq.	max.	PSD
15-10-5	23.65N 160.03E	3	0.20	2e-15	
15-10-6	28.03N 156.46E	3	0.20	1e-15	
15-10-7	33.35N 154.19E	3	0.19	2e-15	
15-10-8	38.44N 155.60E	3	0.18	3e-15	
15-10-9	39.35N 153.57E	4	0.19	9e-16	
15-10-9	-44.48N -114.73E	1	0.15	9e-15	
15-10-10	-49.34N -114.21E	0	0.14	1e-14	
15-10-18	21.98N 139.96E	4	0.21	4e-16	
15-10-19	23.51N 137.24E	4	0.22	4e-16	
15-10-20	40.35N -47.98E	4	0.22	6e-16	
15-10-21	56.70N -39.44E	2	0.17	1e-15	
15-10-21	54.92N -33.56E	4	0.20	2e-15	
15-10-22	59.88N -18.45E	1	0.14	2e-15	
15-10-23	59.74N -25.52E	2	0.16	9e-16	
15-10-26	55.44N -23.51E	0	0.15	6e-15	
15-10-29	57.15N -29.36E	3	0.19	2e-15	
15-11-4	55.44N -23.51E	1	0.15	5e-15	
15-11-11	57.70N 1.43E	1	0.14	6e-15	
15-11-22	49.92N 167.20E	1	0.15	2e-15	
15-11-23	52.37N 172.01E	1	0.14	1e-15	
15-12-1	57.15N -29.36E	4	0.12	4e-15	
15-12-1	59.88N -18.45E	0	0.12	8e-15	
15-12-3	62.98N -25.69E	0	0.12	8e-15	
15-12-7	56.83N -13.73E	1	0.14	3e-15	
15-12-7	56.29N -12.94E	0	0.13	2e-14	
15-12-8	58.25N -19.47E	1	0.16	5e-15	
15-12-9	62.98N -25.69E	0	0.13	9e-15	
15-12-12	42.93N 160.26E	3	0.19	3e-15	
15-12-13	42.93N 160.26E	0	0.13	2e-15	
15-12-21	58.87N -25.63E	0	0.14	8e-15	
15-12-23	59.88N -18.45E	1	0.14	5e-15	
15-12-24	67.81N -3.18E	2	0.16	2e-15	
15-12-25	67.12N 3.20E	2	0.15	3e-15	
15-12-26	47.08N 163.26E	1	0.15	5e-15	
15-12-29	55.57N -19.95E	0	0.13	4e-15	
15-12-30	59.88N -18.45E	0	0.13	7e-15	
15-12-30	49.44N -2.20E	1	0.14	2e-14	
15-12-31	58.87N -25.63E	0	0.14	6e-15	